Mechanisms and impacts of climate tipping elements

Seaver Wang[1], Adrianna Foster[2], Elizabeth A Lenz[3], John D. Kessler[4], Julienne Christine Stroeve[5], Liana O. Anderson[6], Merritt Turetsky[7], Richard Betts[8], Sijia ZOU[9], Wei Liu[10], William R. Boos[11], and Zeke Hausfather[1]

[1]The Breakthrough Institute
[2]National Center for Atmospheric Research
[3]University of Hawaii at Mānoa Sea Grant College Program
[4]University of Rochester
[5]University of Manitoba
[6]National Center for Monitoring and Early Warning of Natural Disasters
[7]University of Colorado Boulder
[8]Met Office
[9]Xiamen University
[10]University of California, Riverside
[11]University of California, Berkeley

November 24, 2022

Abstract

"Climate tipping elements" often refer to large-scale earth systems with the potential to respond nonlinearly to anthropogenic climate change by transitioning towards substantially different long-term states upon passing key thresholds, frequently referred to as "tipping points." In some but not all cases, such changes could produce additional greenhouse gas emissions or radiative forcing that could compound global warming. Improving understanding of tipping elements is important for predicting future climate risks. Here we review mechanisms, predictions, impacts, and knowledge gaps associated with ten notable earth systems proposed to be climate tipping elements. We evaluate which tipping elements are more imminent and whether shifts will likely manifest rapidly or over longer timescales. Some tipping elements are significant to future global climate and will likely affect major ecosystems, climate patterns, and/or carbon cycling within the 21st century. However, assessments under different emissions scenarios indicate a strong potential to reduce or avoid impacts associated with many tipping elements through climate change mitigation. Most tipping elements do not possess the potential for abrupt future change within years, and some proposed tipping elements may not exhibit tipping behavior, rather responding more predictably and directly to the magnitude of forcing. Nevertheless, significant uncertainties remain associated with many tipping elements, highlighting an acute need for further research and modeling to better constrain risks.

Mechanisms and impacts of climate tipping elements

Seaver Wang[1], Adrianna Foster[2], Elizabeth A. Lenz[3], John Kessler[4], Julienne Stroeve[5], Liana O. Anderson[6], Merritt Turetsky[7], Richard Betts[8,9], Sijia Zou[10], Wei Liu[11], William R. Boos[12], Zeke Hausfather[1]

Correspondence to: Seaver Wang (seaver@thebreakthrough.org)

[1]The Breakthrough Institute, Oakland, California 94612, United States

[2]National Center for Atmospheric Research, Climate and Global Dynamics Laboratory, Boulder, United States

[3]University of Hawai‘i Sea Grant College Program, University of Hawai‘i at Mānoa, Honolulu, Hawai‘i 96822, United States

[4]Earth & Environmental Sciences, University of Rochester, Rochester, New York 14627, United States

[5]Centre for Earth Observation Science, University of Manitoba, Winnipeg, Manitoba, Canada

[6]National Center for Monitoring and Early Warning of Natural Disasters (CEMADEN), São José dos Campos, Brazil

[7]Institute of Arctic and Alpine Research, Ecology and Evolutionary Biology Department, University of Colorado Boulder, USA

[8]Met Office, FitzRoy Road, Exeter, EX1 3PB, United Kingdom

[9]Global Systems Institute, University of Exeter, Laver Building, North Park Road, Exeter, EX4 4QE, United Kingdom

[10]State Key Laboratory of Marine Environmental Science & College of Ocean and Earth Sciences, Xiamen University, Xiamen, China

[11]Department of Earth and Planetary Sciences, University of California Riverside, Riverside, California 92521, United States

[12]Department of Earth and Planetary Science, University of California, Berkeley, California, USA

Abstract. “Climate tipping elements” often refer to large-scale earth systems with the potential to respond nonlinearly to anthropogenic climate change by transitioning towards substantially different long-term states upon passing key thresholds, frequently referred to as “tipping points.” In some but not all cases, such changes could produce additional greenhouse gas emissions or radiative forcing that could compound global warming. Improving understanding of tipping elements is important for predicting future climate risks. Here we review mechanisms, predictions, impacts, and knowledge gaps associated with ten notable earth systems proposed to be climate tipping elements. We evaluate which tipping elements are more imminent and whether shifts will likely manifest rapidly or over longer timescales. Some tipping elements are significant to future global climate and will likely affect major ecosystems, climate patterns, and/or carbon cycling within the 21st century. However, assessments under different emissions scenarios indicate a strong potential to reduce or avoid impacts associated with many tipping elements through climate change mitigation. Most tipping elements do not possess the potential for abrupt future change within years, and some proposed tipping elements may not exhibit tipping behavior, rather responding more predictably and directly to the magnitude of forcing. Nevertheless, significant uncertainties remain associated with many tipping elements, highlighting an acute need for further research and modeling to better constrain risks.

Table of Contents

1 Introduction

Global climate change will continue over the 21st century if ongoing human emissions of greenhouse gases and human impacts on terrestrial and marine ecosystems remain unabated (Peters et al., 2020; Raftery, Zimmer, Frierson, Startz, & Liu, 2017). In assessments of impacts associated with climate change, increasing focus is centering around "climate tipping elements" – large-scale earth systems associated with positive feedbacks that could potentially undergo state shifts triggered in response to modest levels of additional warming (Lenton et al., 2008). Research literature and public discourse have alternatively termed such systems "climate tipping points." These tipping elements often carry the potential to drive major carbon cycle feedbacks, changes in regional or global climate patterns, global sea-level rise, and/or ecosystem regime shifts and include mechanisms like Arctic permafrost thaw (Hugelius et al., 2020; M. R. Turetsky et al., 2019) and mass loss from major ice-sheets (Nicholas R. Golledge et al., 2019).

Such climate tipping elements could drive significant additional biodiversity loss, alter the distributions of major ecological biomes, and produce important consequences for human society via climate impacts such as sea-level rise and severe weather events. Some researchers have also proposed that such tipping elements could dynamically interact with one another, producing compounding effects that could ultimately commit the climate system to several degrees of additional warming beyond the level resulting from anthropogenic factors alone (Steffen et al., 2018). Exceeding warming thresholds that could trigger such a cascade has been proposed to carry the potential to alter the long-term climate trajectory of the earth on geologic timescales (Lenton et al., 2019). Apart from the risk of compounding greenhouse gas emissions and radiative forcing, tipping elements carry important implications for the future well-being of human communities and natural ecosystems at local, regional, and global scales. Many tipping mechanisms may also be difficult to halt, reverse, mitigate, or adapt to after state shifts have begun in response to climate perturbations.

Evaluating the projected impacts associated with climate tipping elements and their potential interactions has proven challenging for the earth science community. Most of the most frequently-cited climate tipping elements, such as permafrost carbon release (M. R. Turetsky et al., 2019), disruption of the Atlantic Meridional Overturning Circulation (AMOC) (Lynch-Stieglitz, 2017), degradation and deforestation of the Amazon rainforest (Carlos A. Nobre et al., 2016), and large-scale ecosystem shifts within the northern circumpolar boreal forest (Scheffer, Hirota, Holmgren, Van Nes, & Chapin, 2012), involve highly complex physical and biological systems, competing feedbacks in response to climate change, and large uncertainties. Many of these systems are poorly resolved by the current generation of climate models or omitted altogether, making such mechanisms – alongside carbon cycle feedbacks - an important source of ambiguity regarding climate predictions and global climate sensitivity.

Discussion of “tipping elements” and “tipping points” in the context of earth and climate science research increased sharply in the mid-2000s (Kopp, Shwom, Wagner, & Yuan, 2016) with influential work (Lenton et al., 2008; Lenton & Schellnhuber, 2007) establishing definitions for climate tipping elements and proposing a number of earth systems as examples. Climate tipping elements have since become an increasing focus of research and discussion e.g. (Cai, Lenton, & Lontzek, 2016; Kriegler, Hall, Held, Dawson, & Schellnhuber, 2009; Lenton, 2011, 2012). A recent 2018 research paper assigned elevated risks of triggering a “tipping point cascade” leading to irreversible and significant additional global warming and a “hothouse” climate for warming of >2°C above pre-industrial temperatures (Steffen et al., 2018), and researchers have since continued to develop tipping point cascade theory (Lenton et al., 2019; N Wunderling, Donges, Kurths, & Winkelmann, 2021). At the same time, the concept of climate tipping points has also become a popular topic of discussion within media and political discourse (Extinction Rebellion, 2019; McKibben, 2019).

Definitions of what constitutes a “tipping point” have varied somewhat in the scientific literature regarding whether tipping mechanisms are necessarily irreversible, fast-acting once activated, or triggered at a precise threshold. Inconsistency in terminology can produce confusion, particularly regarding the timescales over which climate mechanisms act.

As outlined by (Kopp et al., 2016), a “tipping point”, as invoked in socioeconomic contexts prior to the term’s adoption by climate researchers, referred to a small change beyond a key threshold triggering networked positive feedbacks and a rapid shift between substantially different system states. In introducing the term in an earth science context, (Lenton et al., 2008) interpreted “climate tipping points” as referring more generally to large state shifts in components of the climate system resulting from relatively small forcings, including examples not involving network-associated feedbacks and/or anticipated to take place over longer time scales of centuries or more. This latter, broader definition has become standard within contemporary discussions of climate tipping points in the earth science community (Lenton et al., 2019; Steffen et al., 2018). Systems categorized as “tipping points” therefore need not refer only to systems capable of rapid change.

Remarking on the varied terminology employed by current literature, (Kopp et al., 2016) identified the need to resolve confusion through improved terminology, proposing the term “tipping elements” to refer to any systems capable of committed nonlinear shifts between states—rapid or gradual—resulting from small changes in forcing. Such committed changes are irreversible on non-geologic timescales, even following subsequent reduction or alleviation in the magnitude of the forcing. Meanwhile, “critical thresholds” refer to the precise level of forcing required to trigger committed system changes. The term “tipping points” would then refer strictly to critical thresholds beyond which tipping elements undergo more rapid “Gladwellian” state shifts, with little temporal lag separating commitment and realization. In contrast, tipping elements might exhibit “non-Gladwellian” behavior, in which the system’s rate of response lags significantly behind the transgression of a critical threshold. The key critical threshold(s) of non-Gladwellian tipping elements, however, would not qualify as “tipping points”, which apply solely to Gladwellian systems. Finally, “non-tipping elements” would refer to systems that respond more predictably or linearly to applied forcing, rather than undergoing committed shifts between system states after crossing key forcing thresholds.

This review largely adopts the convention proposed by (Kopp et al., 2016) to maximize clarity, characterizing systems as “tipping elements” and “non-tipping elements” based on the above definitions. (Kopp et al., 2016) further classified tipping elements as Gladwellian if subject to little time lag between the application of forcing beyond a critical threshold and the system’s response, or categorized as non-Gladwellian if the system instead exhibits considerable lag. We elect instead to characterize tipping elements as abrupt (time lag of less than two decades between threshold transgression and system response), fast (time lag of several decades), or slow (time lag of a century or longer). We employ the term “tipping point” to refer only to critical thresholds for abrupt or fast tipping elements. Further, we use the term “irreversible” to refer only to changes that cannot be halted or returned to the original state on timescales of less than hundreds of years. In contrast, a “reversible” system refers to mechanisms that can be more rapidly reverted to the original state under the right conditions.

Note that the terms “tipping elements” and “non-tipping elements” in and of themselves do not necessarily imply reversible or irreversible changes.

Given the high importance of climate tipping elements for informing future risk assessments and determining optimal societal actions, a firmer understanding of tipping elements with a potentially abrupt or fast response to warming and how they may interact holds the potential to greatly benefit climate discourse. This review seeks to help fulfil this need by synthesizing the latest research on a number of the most frequently-discussed candidate climate tipping elements. Evaluation of the risks posed by climate tipping elements requires considering their timescales of action, climate impacts, and important uncertainties surrounding triggering thresholds and associated factors.

This review explores each of these considerations for ten global and regional candidate climate tipping elements (Table 1) that have featured prominently in broader climate discourse and are commonly thought to present significant global or regional dangers. We also evaluate scientific discussion of the likelihood of a “tipping point cascade” in which multiple tipping elements cumulatively produce substantial further global warming, leveraging a simple climate model (FaIR) (Millar, Nicholls, Friedlingstein, & Allen, 2017; Christopher J. Smith et al., 2017) and finding the additional global mean surface temperature increase over the 21st century with the inclusion of tipping elements to be around 0.76°C (low: 0.55°C, high: 1.00°C) in the high-emissions SSP5-8.5 scenario. We furthermore assess the climate tipping elements that we consider the highest-risk mechanisms in terms of driving more significant, imminent changes to major ecosystems, global carbon cycling, or climate and circulation patterns.

Candidate tipping element	**Classification**	**Timescale of system response**	**Irreversibility**
AMOC weakening/collapse	Tipping element	Slow (century or longer)	Potentially irreversible.
Methane hydrate destabilization	Tipping element	Slow (centuries to millennia)	Irreversible.
Greenland and Antarctic ice-sheet loss	Tipping element	Slow (centuries to millennia)	Irreversible.
Permafrost carbon release	Tipping element	Slow (centuries)	Irreversible.
Boreal forest ecosystem shifts	Tipping element	Potentially fast	Uncertain
Stratocumulus cloud deck evaporation	Tipping element	Potentially abrupt or fast	Irreversible.
Coral reef habitat collapse	Tipping element	Abrupt to fast	Irreversible.

Amazon rainforest dieback	Tipping element	Fast	Irreversible.
Abrupt transitions in S. Asian, African monsoon regime	Nontipping element	Not applicable	Reversible.
Loss of Arctic sea ice	Summer sea ice: Non-tipping element Winter sea ice: Tipping element	Summer sea ice: Abrupt to fast Winter sea ice: Potentially abrupt	Summer sea ice: Reversible. Winter sea ice: Potentially irreversible

Table 1: List of tipping elements discussed in this review, classified as either tipping elements or nontipping elements based on the terminology of (Kopp et al., 2016)**. The timescale of system-wide response to any tipping behavior is further classified as abrupt (taking place within a couple decades with no time lag), fast (several decades), or slow (longer than a century). Finally, impacts are classified as reversible (reversion to original system state within centuries possible upon applying opposite forcing), or irreversible (reversion to original system state requires centuries or longer, and/or different opposite forcing of a significantly larger magnitude than the original change in forcing applied to achieve the altered system state).**

2 Global candidate tipping elements

2.1 Slowdown or collapse of the Atlantic Meridional Overturning Circulation (AMOC)

2.1.1 Background

The global Meridional Overturning Circulation (MOC), also referred to as the thermohaline circulation (THC), is a critical component of the climate system. In the Atlantic basin, the Atlantic MOC (AMOC) is characterized by warm, saline waters that move northward in the upper ocean by the Gulf Stream and the North Atlantic Current (Buckley & Marshall, 2016). Cooling as they travel north along these pathways, these waters increase in density and lose their buoyancy, sinking once they reach a critical density threshold and becoming North Atlantic Deep Waters (NADW), which then return southward in the deep ocean (W.S. Broecker, 1991).

Due to its circulation structure, the AMOC impacts the northern hemisphere weather and climate by redistributing heat between the low and high latitudes (Bryden et al., 2020; Wunsch, 2005). In addition, the AMOC is shown to play a key role in the uptake and export of anthropogenic carbon (Takahashi et al., 2009), and influence long-term global patterns of ocean circulation and carbon cycling, with important consequences for global warming and oceanic primary productivity (Menviel, Timmermann, Mouchet, & Timm, 2008; Muglia, Skinner, & Schmittner, 2018).

On interannual timescales, the AMOC is driven by large-scale wind patterns across the North Atlantic according to observations at 26.5°N (Moat et al., 2020; C. D. Roberts et al., 2013) and model simulations (Arne Biastoch, Böning, Getzlaff, Molines, & Madec, 2008; Zhao & Johns, 2014). Over decadal timescales and beyond, temperature and salinity influences exert a strong influence over this overturning circulation (Arne Biastoch et al., 2008; Buckley, Ferreira, Campin, Marshall, & Tulloch, 2012; Buckley & Marshall, 2016).

Ongoing climate change has sparked concerns that this important feature of ocean circulation could be disrupted by rising water temperatures and melting ice-sheets. A slowdown or shutdown of the AMOC system would significantly affect regional and global climate patterns. Paleoclimate evidence has identified weakening of the AMOC as an important potential driver of multiple large, rapid shifts in past climate, including fast or abrupt changes occurring on timescales as short as a few decades (Alley, Anandakrishnan, & Jung, 2001; P. U. Clark et al., 2001). The impacts of past AMOC shifts affected climate globally, significantly altering tropical rainfall patterns and causing heat redistribution between the northern and southern hemispheres (Masson-Delmotte et al., 2013). Changes to the overturning circulation could also affect the ocean's strength as a heat and carbon sink.

The abovementioned paleoclimate changes indicate a potentially bi-stable AMOC during past climates. Driven by the salt-advection feedback (Stommel, 1961), the AMOC could switch between "on" and "off" states under natural perturbations such as deglacial meltwater pulses when the ocean system passes certain tipping points. These sudden AMOC changes have been suggested as one of the main mechanisms explaining the climate changes during the Dansgaard-Oeschger and Heinrich events (Wallace S Broecker, Peteet, & Rind, 1985; Peter U. Clark, Pisias, Stocker, & Weaver, 2002; Stefan Rahmstorf, 2002). Moreover, based on an AMOC stability indicator (W. Liu & Liu, 2013; S Rahmstorf, 1996), analyses of modern observations suggest that the current AMOC resides in a bi-stable regime. The circulation may be at risk of eventual collapse under future anthropogenic warming, as the possibility of AMOC collapse could be downplayed currently by most coupled climate models due largely to a ubiquitous model bias towards AMOC stability (W. Liu, Liu, & Brady, 2014; W. Liu, Xie, Liu, & Zhu, 2017).

Increasing temperatures in combination with heightened meltwater fluxes from the Greenland Ice-sheet (GIS) could warm and freshen surface waters to the point of slowing or even halting the formation of NADW (P. Bakker et al., 2016). As the twin factors of temperature and salinity play key roles in the buoyancy loss process responsible for driving the overturning circulation, warming and freshening of surface waters may inhibit the sinking process that drives NADW formation from taking place. Consequently, the climate and oceanographic research communities are devoting considerable attention towards determining whether the AMOC is currently weakening and assessing its risk of weakening or collapse in response to future warming beyond a certain threshold.

2.1.2 Evidence for a weakening AMOC warrants concern given uncertainties in response to climate change

Assessments of whether the AMOC is weakening or has weakened since the pre-industrial era are complicated by a relative paucity of direct measurements of the AMOC's strength. The first continuous, basin-wide observations of the AMOC only became available with the installation of the RAPID-MOCHA instrument array in 2004 (Cunningham et al., 2007). A new monitoring array, OSNAP, was installed in the subpolar North Atlantic at approximately 52°N-60°N in 2014 (M. Lozier et al., 2017). At 34.5°S, moored arrays SAMBA were deployed in 2009-2010 and since 2013 to understand the overturning variability in the South Atlantic (Kersalé et al., 2018). Consequently, direct measurements of the overturning circulation have only been available for 15 years. To assess the impact of anthropogenic climate forcing upon AMOC strength over the past century and beyond, scientists have leveraged proxy data to estimate historical changes in the AMOC.

Several proxy-based studies have presented evidence that the AMOC's strength has fallen over time since the onset of the industrial age. One analysis utilized silt grain size in sediment cores taken offshore of North Carolina as a proxy for bottom current flow speed and found a significant weakening trend distinct from the pattern of AMOC strength over the past 1,500 years (Thornalley et al., 2018). Uncertainty associated with precisely dating the age of silt grains as well as potential sediment mixing from physical or biological activity represent potential sources of error, although estimates of AMOC strength from sediment cores do demonstrate good agreement with temperature-based reconstructions. Furthermore, this study suggests that AMOC weakening may have begun early in – if not prior to – the industrial era, and the authors note that the degree to which potential weakening is associated with anthropogenic forcing remains an open question.

A 20th century AMOC decline is consistent with a number of other proxy records, including the North Atlantic SST and subsurface temperature, and $\delta^{18}O$ in benthic foraminifera (Stefan Rahmstorf et al., 2015). The decline was likely accelerated around 1970, which was followed by a partial recovery in the 1990s before rapidly accelerating again in the mid-2000s (Caesar, McCarthy, Thornalley, Cahill, & Rahmstorf, 2021; Caesar, Rahmstorf, Robinson, Feulner, & Saba, 2018; Stefan Rahmstorf et al., 2015). A more recent publication used 11 proxies, including reconstructed SST, $\delta^{18}O$, and $\delta^{15}N$ and found a consistent rapid decline of AMOC in the mid-20th century (Caesar et al., 2021).

Other studies seek to reconstruct AMOC strength using hydrographic sections, satellite altimetry measurements and reanalysis products. With five transatlantic sections along 25°N, (Bryden, Longworth, & Cunningham, 2005) reported a decline of the AMOC by 30% between 1957 and 2004. At similar latitude, (Frajka-Williams, 2015) constructed AMOC transport strength back to the mid-1990s using altimetry and cable measurements, and found a slight decrease of the AMOC. The AMOC decline was also simulated at subpolar latitudes in an ocean reanalysis product (Laura C Jackson, Peterson, Roberts, & Wood, 2016). Contrastingly, reconstructed AMOC at a mid-latitude was found to exhibit a slight increase since the early 1990s (Willis, 2010). The discrepancy is likely attributed to the sensitivity of reconstructed AMOC to different calculation and assimilation procedures (Karspeck et al., 2017), as well as the latitudinal dependence of AMOC strength variability (Arne Biastoch et al., 2008; Bingham, Hughes, Roussenov, & Williams, 2007). Estimates of the AMOC state since the 1990s using model-based reanalysis and data from hydrographic sections also support a strong role of interannual to decadal variability in driving the lack of a distinct trend in AMOC strength over recent decades (Y. Fu, Li, Karstensen, & Wang, 2020; L C Jackson et al., 2019).

While observational records are too short to conclude on AMOC long-term change, they so far have revealed a number of unprecedented aspects of the AMOC and its variability, with implications on how we interpret the hydrographic and proxy-based reconstructions as well as model simulations. Observations from the RAPID monitoring array have revealed significant AMOC variability from daily to inter-decadal on interannual, interannual, and decadal time scales, with a range of ~4-35 Sverdrups (1 Sv = 1 million cubic meters per second) over one year (Srokosz & Bryden, 2015), implying the inadequacy of inferring AMOC changes from hydrographic snapshots. Additionally, the first four years of OSNAP observations have suggested that the AMOC strength is mostly explained by water mass transformation in ocean basins east of Greenland, instead of in the Labrador Sea west of Greenland as suggested by historical paradigm (F. Li et al., 2021; M. S. Lozier et al., 2019). The minimal contribution of the Labrador Sea to the AMOC is further attributed to the compensating effects of cold and fresh anomalies on density in the basin's boundary current (Pickart & Spall, 2007; S. Zou, Lozier, Li, Abernathey, & Jackson, 2020), highlighting the equally important role of heat and freshwater flux in determining the AMOC strength.

These results suggest that proxy records of either temperature or salinity alone may not be adequate to reconstruct overturning variability and highlight the importance for climate models to accurately simulate water mass transformation/formation process, including when, where and how deep waters are produced, mixed, and exported (Heuzé, 2017; M. S. Lozier et al., 2019). Yet at the same time, the strong agreement across such models that AMOC strength is expected to decline under continuing emissions scenarios (Collins et al., 2019; Menary & Wood, 2018) emphasizes that continued efforts to refine existing models and re-assess risks would be prudent.

Whereas assessments of whether the AMOC is currently weakening remain subject to some uncertainty, the research community stands largely in agreement that a complete collapse of the AMOC in the near term is a low-probability event. The IPCC's Special Report on Oceans and the Cryosphere in a Changing Climate concluded that an AMOC collapse in the 21st century was "very unlikely," with only one of 27 models producing an AMOC collapse by 2100, and even then only under the worst-case RCP8.5 high-emissions scenario (Collins et al., 2019). Modeling results from a 2016 study concluded that the likelihood of an AMOC collapse under RCP8.5 was 44% by 2290-2300, while AMOC collapse was averted altogether under the RCP4.5 pathway (~2.5°C warming in 2100) (P. Bakker et al., 2016). However, current coupled climate models exhibit biases in surface ocean climatology that favor greater AMOC stability (W. Liu et al., 2014). A modeling analysis correcting for these biases and assuming a CO_2 doubling approximately between the RCP4.5 and RCP6.0 scenarios produced an AMOC collapse 300 years after the CO_2 perturbation (W. Liu et al., 2017). Still, a need remains for improved model physics to more realistically investigate the response of AMOC to different climate change scenarios.

Overall, the possibility that the overturning circulation is currently weakening and may weaken further with continuing warming is sufficiently backed by recent research to justify the degree of past and ongoing attention devoted to this potential climate tipping element.

2.1.3 Regional impacts of a weakened AMOC

The consequences of an AMOC collapse would be of substantial magnitude and global in scale. A shutdown of the overturning circulation has been modeled to cause significant North Atlantic cooling (3-8°C), South Atlantic warming (0-3°C), reduced North Atlantic rainfall (-0.5 to -2 mm/d), and a southward shift of the tropical rain belt (by ~10° latitude) (L. C. Jackson et al., 2015) and western North Atlantic sea level rise. Considerable paleoclimate evidence supports the existence of linkages between AMOC weakening and Northern Hemisphere cooling that would be accompanied by Southern Hemisphere warming, as reviewed by (Lynch-Stieglitz, 2017).

In contrast to total collapse, however, weakening of the AMOC results in more moderate impacts. Modeling a freshwater-induced reduction in AMOC strength from 16 to 9 Sverdrups yielded North Atlantic subpolar gyre cooling of 2°C (Haarsma, Selten, & Drijfhout, 2015). A modeling exercise isolating the climate impacts resulting from a weakening AMOC under future warming produced similar findings, including North Atlantic cooling, shifts in rainfall and atmospheric circulation, and deceleration of Arctic sea ice loss (W. Liu, Fedorov, Xie, & Hu, 2020). These analyses modeled weakening of the AMOC somewhat in excess of what IPCC CMIP5 models have predicted: declines on the order of –5.5 ± 2.7 Sv by 2100 under the high-end RCP8.5 scenario (Figure 1) (Collins et al., 2019). Multimodel analyses of new CMIP6 models, however, has found a potential for stronger decreases in the AMOC over the 21st century (M. J. Roberts et al., 2020; Weijer, Cheng, Garuba, Hu, & Nadiga, 2020), with mean declines of 6-8 Sv by 2100 and, for the upper-end RCP8.5 scenario, a range encompassing a 6-11 Sv potential reduction in strength (Weijer et al., 2020). At the same time, concern exists that current climate models overestimate the stability of the AMOC in response to warming, while computational and cost limitations prevent assessment of AMOC stability over millennial timescales using high-resolution ocean models, as reviewed by (Weijer et al., 2019).

Some researchers have suggested that the response of the AMOC to climate forcing may depend not only on the transgression of a freshwater forcing threshold but also upon the duration of such an overshoot. (Ritchie, Clarke, Cox, & Huntingford, 2021) utilized a simplistic model of the AMOC to illustrate how a sufficiently short overshoot of a climate threshold followed by a rapid reversion in forcing strength to below the critical level could allow the system to return to the original state, while a longer overshoot would lead to committed, long-term system state change. Studies using coupled climate models have shown this varying pattern of recovery for the AMOC in response to freshwater hosing perturbations of different strengths and lengths of time (Alkhayuon, Ashwin, Jackson, Quinn, & Wood, 2019; L C Jackson & Wood, 2018).

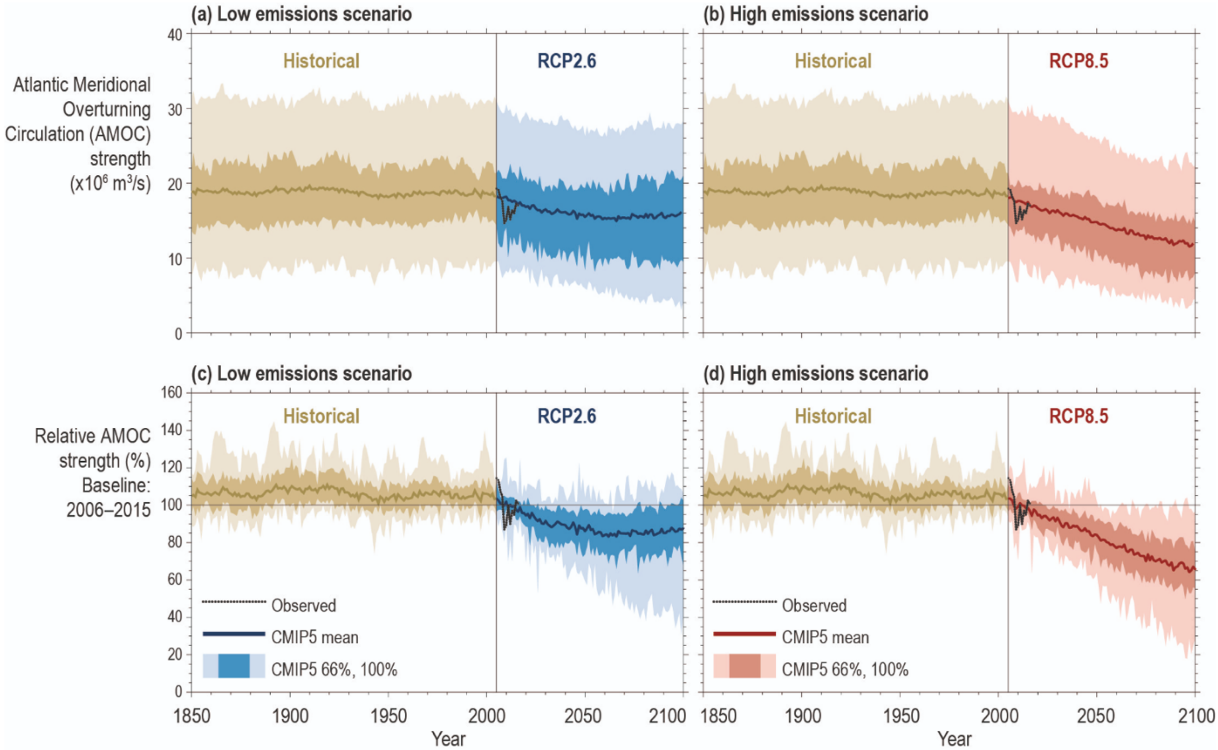

Figure 1: Model simulations of AMOC strength at 26°N under (a, c) low-emissions (RCP2.6) and (b, d) worst-case emissions (RCP8.5) scenarios. Dark black line indicates observational estimates determined from the RAPID instrument array at 26°N. Gray, blue, and red lines show multi-model ensemble mean values (27 models in total). Darker and lighter shading show 66% and 100% ranges for model runs, respectively. Model results are expressed in estimated AMOC water mass transport rates (a, b) and in relative percentage change compared to the 2006-2015 baseline (c, d). Figure reproduced with permission, originally published as Figure 6.8 in the IPCC SROCC Ch 6 (Collins et al., 2019)**.**

Changes in the AMOC may also alter storm tracks and intensities in the N. Atlantic (Gastineau, L'Hévéder, Codron, & Frankignoul, 2016) as well as Arctic sea ice patterns (Delworth & Zeng, 2016; Yeager, Karspeck, & Danabasoglu, 2015). Evidence of sea-level changes along parts of the eastern coast of North America linked to AMOC variability (Ezer, Atkinson, Corlett, & Blanco, 2013; Goddard, Yin, Griffies, & Zhang, 2015; Little et al., 2019) also suggests that a weakening AMOC may have implications for adaptation to sea-level rise in this region. Climate modelling supports a potential for AMOC weakening to drive accelerated sea-level rise in the North Atlantic basin (Levermann, Griesel, Hofmann, Montoya, & Rahmstorf, 2005; Yin & Goddard, 2013; Yin, Schlesinger, & Stouffer, 2009) with a follow-up study predicting meaningful regional sea-level impacts even under lower emissions scenarios (C F Schleussner, Frieler, Meinshausen, Yin, & Levermann, 2011). Some of the most acute AMOC-induced impacts for human communities include substantial potential reductions in rainfall in the Sahel region, negatively impacting crop yields (Defrance et al., 2017). Other global-scale interactions with the Pacific Ocean and the Antarctic remain uncertain but are the subject of continuing research (Collins et al., 2019).

Slowdown of the AMOC is also theorized to represent a positive feedback on the carbon cycle due to reduced carbon uptake and possible wetland methane-related feedbacks (Luke A. Parsons, Yin, Overpeck, Stouffer, & Malyshev, 2014). The possibility also exists that an AMOC slowdown may interact with other potential tipping elements within the climate system, such as the potential for Amazon forest dieback or the acceleration of ice loss from the West Antarctic Ice-sheet (Steffen et al., 2018).

Finally, defining particular critical temperature thresholds expected to contribute to committed weakening of the overturning circulation also represents a challenge (Weijer et al., 2019). (O. Hoegh-Guldberg et al., 2018) determines a higher likelihood of more intense weakening for >2°C of warming based on model predictions. Given that substantial committed loss of the Greenland Ice-sheet is also more likely to occur than not beyond a 2°C warming threshold (Pattyn et al., 2018), with important implications for buoyancy dynamics in deep water formation regions, the IPCC's assessment of a 2°C threshold seems a plausible lower bound above which the risks of significant weakening of the AMOC increase. However, the current ability of models to accurately represent the AMOC and predict its response to climate change remains low, leaving the proximity of today's AMOC to potential critical thresholds uncertain (Weijer et al., 2019). Further improvements to our understanding of the AMOC and additional advances in model complexity and computing power will prove crucial for better constraining potential long-term outcomes.

2.2 Large-scale release of methane from destabilization of marine methane hydrate deposits

2.2.1 Background

Methane (CH_4) is a potent greenhouse gas whose global warming potential has been revised slightly upwards over the last few decades to a current figure of 34 CO_2e (CO_2 equivalents) over a 100-year time horizon, and 86 CO_2e based on a 20-year time horizon (IPCC, 2013). The majority of atmospheric methane is biological in origin, often the product of methane-emitting microorganisms undergoing respiration under low or no-oxygen conditions (Dean et al., 2018). Wetlands, inundated rice fields, and gases released from the gastrointestinal tract of ruminant herbivores like cows, for instance, all represent significant fluxes of methane globally. The remainder of atmospheric methane is of fossil origin, released by activities such as oil, gas, and coal extraction and production.

Estimates of methane fluxes exhibit some range of uncertainty, with the range of estimates for methane emissions from particular sources further varying between “top-down” and “bottom-up” methods (Table 2). “Bottom-up” methods estimate methane fluxes associated with each category of sources using small-scale measurement data, emissions records, and land surface modeling. In contrast, “top-down” methods use observed atmospheric methane measurements and inverse-modelling techniques to attribute fluxes to different sources (M Saunois et al., 2020). New research continues to update and refine the global methane budget, with one recent study finding that a proportion of methane emissions previously attributed to natural geological sources are actually of anthropogenic fossil origin, thus reapportioning estimated methane fluxes from these two sources (Hmiel et al., 2020). Over the coming century and beyond, the global methane budget is anticipated to evolve further in response to climate change and human activities, with a major concern being that methane emissions from sources that are currently minor or negligible could increase significantly due to warming.

Methane source	**Emission rate (Mt CH_4 per year)**			
	Kirschke et al. 2013 (bottom-up estimates)	**Schwietzke et al. 2016**	**Saunois et al. 2020 (full range of reported top-down and bottom-up estimates, 2000-2017)**	**Hmiel et al. 2020**
Wetlands	217 (177-284)		100-217	
Agriculture*	200 (187-224)		178-246 (includes landfills and waste)	
Fossil fuels*	96 (85-105)	145±23	71-164	177±37
Geological	54 (17-97)	51±20	18-65	1.6
Freshwater	40 (8-73)		117-212	
Biomass burning*	35 (32-39)		22-46	
Wild animals	15 (15-15)		1-3	
Termites	11 (2-22)		3-15	
Methane hydrates	6 **(IPCC AR5 estimate for methane hydrate release.)**		0	
Wildfires*	3 (1-5)		1-5 (estimate from (Marielle Saunois et al., 2016), possible double-counting with biomass burning)	

Permafrost	1 (0-1)		0-1	

Table 2: Current estimates for sources of atmospheric methane in Mt CH_4 yr^{-1} based on (Kirschke et al., 2013)**,** (Schwietzke et al., 2016)**,** (M Saunois et al., 2020)**, and** (Hmiel et al., 2020)**. Asterisks denote anthropogenic methane sources, including wildfires. When combining a range of multiple estimates, uncertainties are not shown.**

Significant deposits of methane are currently sequestered in ocean sediments as methane clathrates or methane hydrates. This solid, ice-like form of methane is stable at high pressures, low temperatures, and high methane concentration conditions. Typically, water depths of greater than 500 m are necessary to establish the pressure/temperature conditions necessary for marine methane hydrates to form in sediment; yet in colder Arctic waters, only around 300 m in depth is required (Carolyn D. Ruppel & Kessler, 2017). The methane encapsulated in hydrate deposits is largely produced from microbial decomposition of organic matter in ocean sediments (Claypool & Kvenvolden, 1983) or thermogenic processes (Sassen et al., 1999). Estimates of the global abundance of methane sequestered in hydrates places them as one of the largest organic carbon reservoirs on Earth (Carolyn D. Ruppel & Kessler, 2017).

Whether fast or more gradual, the destabilization of even a fraction of the global marine hydrate reservoir could hold implications for global climate through the release of a greenhouse gas and a reactive carbon species to the environment. As a result of being positioned at shallower depths and the significant warming currently experienced at high latitudes, Arctic methane hydrate deposits are thought to be the most vulnerable pool of marine methane hydrates to warming-induced thaw. With the Arctic Ocean containing a large percentage of the planet's total continental shelf area, a significant inventory of marine hydrate (~116 Gt C) exists across this basin at relatively modest depths (K. Kretschmer, Biastoch, Rüpke, & Burwicz, 2015). Consequently, the scientific community is actively assessing the potential for a large-scale thaw and release of methane hydrate from the Arctic seafloor should increasing sediment temperatures destabilize hydrate formations.

In the absence of widespread sediment failure on the continental slope, methane release from decomposing hydrates will likely manifest as a gradual flux of methane from the seafloor over long timeframes, due to the dynamics of sediment column warming (D Archer, 2015; David Archer, Buffett, & Brovkin, 2009; Carolyn D. Ruppel & Kessler, 2017; Carolyn D Ruppel, 2011). Slow release carries important implications for warming. First, assuming all seafloor-released methane is emitted to the atmosphere, methane's rapid decay (~10 years) to carbon dioxide in the atmosphere would mean that this longer-term release would be effectively converted to a far less-powerful greenhouse gas on a molecule-for-molecule basis. Furthermore, several mechanisms in both sediment and overlying waters will act as powerful sinks that consume rising methane from thawing deposits, preventing much of the destabilized methane from reaching the atmosphere (Figure 2). Currently, the contribution of methane hydrates to global atmospheric methane fluxes is considered negligible (M Saunois et al., 2020). Consequently, the destabilization of methane hydrates in marine sediments represents a low threat for atmospheric methane emissions, although it remains a climate tipping element able of contributing significant amounts of reactant carbon into Earth systems.

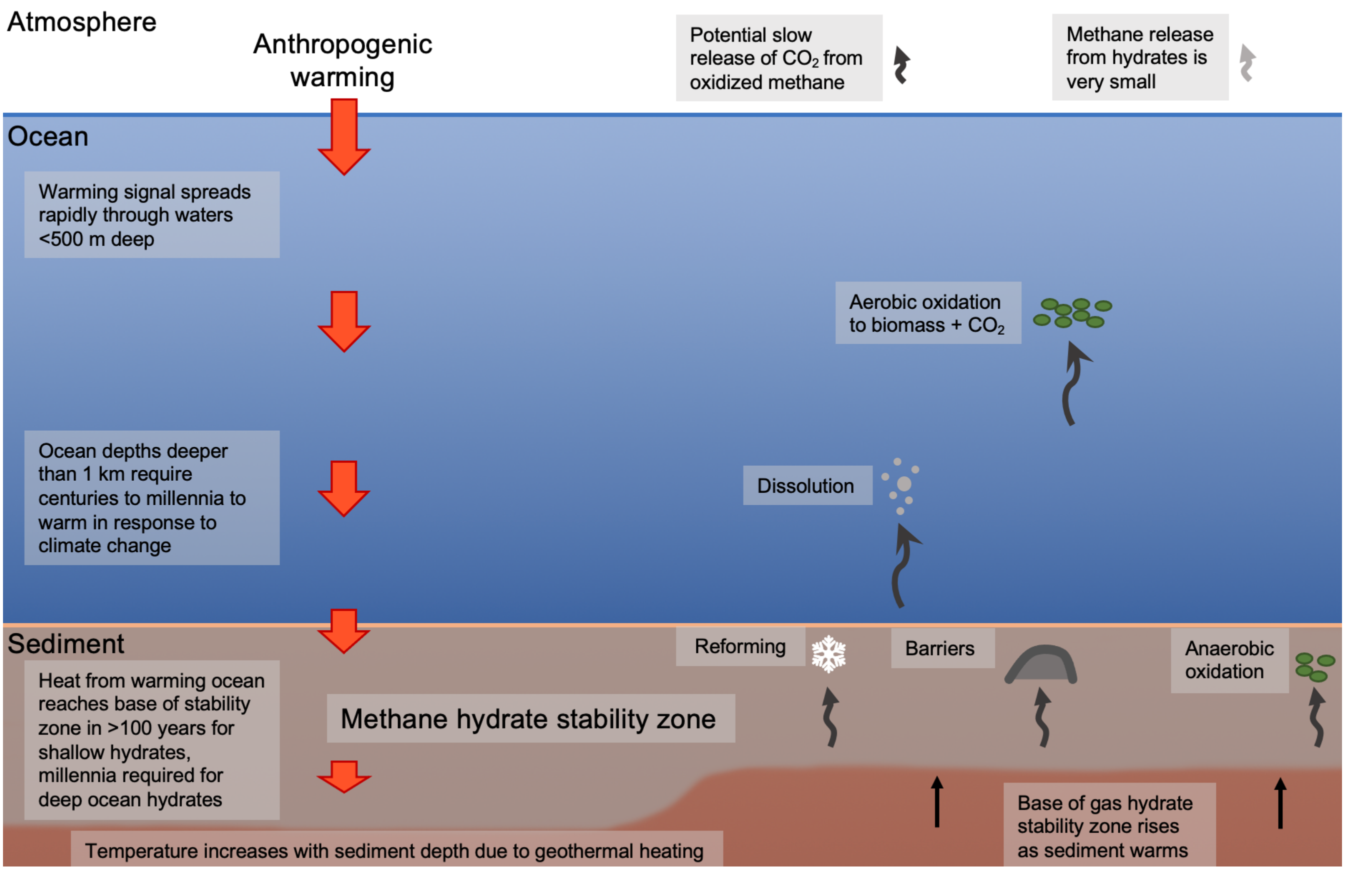

Figure 2: Schematic diagram of marine and sedimentary sources and sinks for methane and carbon dioxide derived from gas hydrate deposits, along with timescales required for warming to reach effective depth within methane hydrate stability zone.

Carbon stock	**Total carbon (Gt C)**
Methane hydrates (2017 estimate)	1800 Gt C (Carolyn D. Ruppel & Kessler, 2017)
Methane hydrates (1988 estimate)	11000 Gt C (Kvenvolden, 1988)
Fossil fuels	5000 Gt C
Soil carbon	1400 Gt C

Dissolved marine carbon	980 Gt C
Land biomass	830 Gt C
Peat	500 Gt C
Other	67 Gt C

Table 3: Comparison between updated and initial estimates of carbon stored within methane hydrates, relative to other major stocks of organic carbon on Earth. Adapted from charts published in (Carolyn D. Ruppel & Kessler, 2017)

2.2.2 Methane hydrate dissociation would act as a "slow tipping point" with correspondingly gradual impacts

Estimates of total carbon contained within marine methane hydrates globally vary widely, but have largely fallen since the 1990s as important assumptions used in early estimates were overturned by field sampling (Table 3) (Carolyn D. Ruppel & Kessler, 2017). Calculations range between ~1100 Gt C (K. Kretschmer et al., 2015) and ~12,400 Gt C (G. R. Dickens, 2011), with numerous studies converging on values towards the lower end of this range (<2000 Gt C) (David Archer et al., 2009; Boswell & Collett, 2011; Milkov, 2004; Piñero, Marquardt, Hensen, Haeckel, & Wallmann, 2013). As much of the methane hydrate carbon pool globally is located at deeper depths in bottom ocean sediments with higher pressures, large amounts of this carbon (95%) are not considered to be vulnerable to dissociation or release due to projected ocean warming (K. Kretschmer et al., 2015; Carolyn D Ruppel, 2011). However, methane clathrates can co-occur with free methane trapped below this largely impermeable barrier, thus in regions vulnerable to dissociation, the quantity of free methane at risk of release may be up to two-thirds higher than the local methane hydrate pool (Hornbach, Saffer, & Holbrook, 2004).

A significant time lag separates atmospheric warming due to climate change and the much longer timescales required for transport and diffusion of heat anomalies into the ocean and sediment. As sediment warming is required for methane hydrate instability, dissociation may not be initiated until centuries to millennia after the requisite warming spike (D Archer, 2015; David Archer et al., 2009; K. Kretschmer et al., 2015; Carolyn D Ruppel, 2011). For deep ocean sediments, tens of millennia might be required for the methane hydrate zone to begin appreciably warming, let alone for hydrate to begin dissociating (David Archer et al., 2009; Carolyn D Ruppel, 2011). This factor does not preclude eventual significant release of carbon from methane hydrate, but does mean that this climate feedback occurs with a very substantial delay between commitment and realization.

A number of further factors likely ensure that future methane release from hydrate deposits will occur as a gradual emission rather than an abrupt spike. In particular, a number of physical mechanisms will act to greatly slow down the rate of methane escaping from sediments into the water column. Hydrate dissociation is an endothermic process (heat is consumed by the thaw process), limiting the rate at which methane hydrates respond to increased heat flux (Circone, Kirby, & Stern, 2005;

Gupta, Lachance, Sloan, & Koh, 2008; C D Ruppel & Waite, 2020). Hydrate deposits are also generally expected to thaw first at deeper depths in the sediment column. This is because the stability of frozen methane hydrates is greatest at the seafloor, with hydrates becoming less stable as sediment temperature increases with depth due to geothermal heating from below (Zatsepina & Buffett, 1998). Consequently, a “cold trap” exists for destabilized methane migrating upwards to shallower, colder depths in sediment, encouraging reformation of methane hydrate (National Research Council, 2013). Field observations have also found marked declines in methane seepage from marine sediments during the cold season, an effect that may be leading to overestimation of methane release rates calculated only from warm season measurements (Ferré et al., 2020).

Physical barriers can significantly slow or halt the release of methane from sediments. These include factors such as low permeability and barriers to upward migration. At the same time, other physical features such as faults may facilitate the delivery of released methane to the water column (J D Kessler, Reeburgh, Southon, & Varela, 2005). Abrupt physical events such as submarine landslides could also destabilize hydrate deposits, although even the large Storegga landslide (8150 years ago, 5 Gt C released) did not release sufficient quantities of methane to measurably affect climate (D Archer, 2007). Such considerations as well as significant limitations of the bubble volume threshold and other important parameters considerably limit the ability of models to confidently predict methane release in response to warming.

Current models also do not explicitly account for a number of chemical, physical, and biological methane sinks within both sediments and ocean waters. Microbial activity within the sediment column can consume a significant percentage of escaping methane gas (variable by site but up to 80-90%) (Reeburgh, 2007). Even for methane that escapes methane oxidation in sediments, a significant fraction dissolves in the overlying waters and is oxidized prior to atmospheric emission. (McGinnis, Greinert, Artemov, Beaubien, & Wüest, 2006) determined that for 50% of methane within a bubble escaping from the seafloor to reach the atmosphere from a depth of 100 m, the bubble must be two centimeters in diameter. However, since methane hydrates are stable below approximately 500 m depth, even larger bubbles are required to drive substantial methane fluxes. (Binbin Wang, Jun, Socolofsky, DiMarco, & Kessler, 2020) measured bubble size distributions at various altitudes above the seafloor and determined that most bubble diameters range from 2-6 mm and effectively dissolve prior to any atmospheric emission. Following this physical sink, microbial activity within the water column will also act to oxidize escaping methane to CO_2 (Chan et al., 2019; Elliott, Maltrud, Reagan, Moridis, & Cameron-Smith, 2011; John D. Kessler et al., 2011; Leonte, Ruppel, Ruiz-Angulo, & Kessler, 2020; Mau, Blees, Helmke, Niemann, & Damm, 2013; National Research Council, 2013; Valentine, Blanton, Reeburgh, & Kastner, 2001). Both the sediment and water column biological sinks would be expected to intensify slightly with increasing temperatures due to the favorable effect of temperature upon microbial metabolic processes, acting as a negative feedback. Transformation of methane to CO_2 represents a tradeoff, where methane emitted to the atmosphere results in larger immediate climate impacts, whereas CO_2 dissolved in seawater may contribute to ocean acidification (Garcia-Tigreros et al., 2021). CO_2 derived from oxidized hydrate methane could be emitted to the atmosphere as a longer-lived greenhouse gas with a lower radiative forcing than methane, however, this would require this CO_2 to also escape the ocean's biological pump. We further highlight that the atmospheric emission of CO_2 produced from the oxidation of hydrate methane is a modeled effect that has not been directly measured. Overall, the combined effect of such sinks greatly limits the potential magnitude and rate of any atmospheric release of methane from dissociating hydrate deposits.

A number of older modeling analyses had produced much larger climate impacts resulting from methane hydrate release under warming scenarios. However, much of this earlier work (Gornitz & Fung, 1994; Harvey & Huang, 1995) significantly overestimated the available methane hydrate in marine sediments and/or oversimplified the propagation of heat through the ocean and sediment, while omitting important methane sinks in sediment and the water column, as reviewed by (National Research Council, 2013; Carolyn D. Ruppel & Kessler, 2017).

More recent modeling analyses suggest that the climate impacts of methane hydrate release would be gradual and moderate if not minimal. For a 3°C level of warming, one study calculated, over multiple millennia, an additional 0.4-0.5°C temperature increase as a result of methane hydrate destabilization of 940 Gt C (David Archer et al., 2009). However, the quantity of carbon released in response to warming is overwhelmingly driven by choice of key parameters governing the migration of methane through seafloor sediment and the water column, making modeled climate impacts highly uncertain. Higher thresholds simulating more limited passage of methane through sediment led to insignificant quantities of released carbon for negligible climate impact. Further, for a total anthropogenic CO_2 forcing of 1,000 Gt C resulting in 2°C of warming, methane hydrate dissolution was modeled to contribute less than an additional 0.5°C warming even assuming that all dissociated methane reached the atmosphere. In contrast, an assumption of total conversion of escaping methane to CO_2 only extended the duration of 2°C warming, without producing an additional temperature increase (David Archer et al., 2009). Due to the slow thermodynamics of sediment heating and physical and biological sinks, overall methane hydrate emissions thus represent gradual long-term additions of methane and carbon dioxide to seawater and possibly the atmosphere as opposed to threatening an abrupt or fast methane release.

Present-day and near-future methane hydrate fluxes are not thought to represent climatically significant greenhouse forcings. Contentious estimates of high rates of methane emissions from hydrate dissociation on the East Siberian Arctic Shelf (Shakhova et al., 2014) have been revised substantially downwards by numerous subsequent studies (Berchet et al., 2016; Thornton, Geibel, Crill, Humborg, & Mörth, 2016; Thornton et al., 2020; Tohjima et al., 2020). Present-day marine methane release from Arctic hydrate dissociation is likely primarily of natural origin, resulting from the pressure decrease associated with isostatic uplift following the last glacial maximum, rather than a response to anthropogenic forcing (Wallmann et al., 2018).

A near-term modeling study (K. Kretschmer et al., 2015) determined that relative to current *annual* anthropogenic methane emissions of 335 Mt C per year, warming caused by doubling CO_2 concentrations over the 21st century would release a *total* quantity of just 473 Mt of methane from destabilized hydrates over a 100-year period. A more recent modeling analysis similarly concluded that the quantity of methane within shallower, more vulnerable hydrate deposits and the likely rates of dissociation were too low to impart a significant climate impact within the next few centuries under current climate change (Mestdagh, Poort, & De Batist, 2017). Although the last deglaciation does not represent a perfect analogue to contemporary warming, analysis of atmospheric concentrations from 18,000 – 8,000 years ago suggest that release of ancient methane from permafrost and seafloor hydrates in response to past warming events may have been small (<19 Mt CH_4/yr) (Dyonisius et al., 2020; Petrenko et al., 2017).

In conclusion, while levels of warming exist beyond which large quantities of methane in hydrate deposits may eventually become destabilized, numerous physical, thermodynamic, chemical, and biological factors combine to substantially limit the rate at which this methane might escape to the atmosphere. For more moderate warming of ~2°C, methane hydrates might well exert a negligible overall impact on atmospheric temperatures. Methane hydrate dissociation would additionally take place on extremely long timescales of millennia, rather than over abrupt or fast timescales that would produce an acute warming spike. This review primarily focuses on the impact of radiative forcing caused by the release of methane and CO_2 to the atmosphere associated with dissociating methane hydrates. It does not comment on other consequences of hydrate dissociation such as ocean acidification or the decrease in the ocean's ability to be a sink of anthropogenic CO_2, both of which are caused by the increase in dissolved CO_2 concentration from methane oxidation. While these processes are less studied than the direct emission of methane to the atmosphere, some preliminary investigations show that it could be important over longer timescales similar to the invasion of anthropogenic CO_2 into surface seawater (A Biastoch et al., 2011; Boudreau, Luo, Meysman, Middelburg, & Dickens, 2015; Garcia-Tigreros & Kessler, 2018; Garcia-Tigreros et al., 2021). Consequently, in relation to other tipping elements covered within this review, marine methane hydrates represent a relatively minor climate threat.

2.3 Multi-meter sea-level rise from loss of major Greenland and Antarctic ice-sheets

2.3.1 Background

Currently, the majority of sea-level rise (~3.5 mm/yr) (Global Sea Level Budget Group, 2018) is driven by thermal expansion of ocean waters in response to rising temperatures (~1.40 mm/yr) as well as water inputs from melting glaciers (~0.61 mm/yr) outside of the major Greenland and Antarctic ice-sheets. By contrast, the Greenland Ice-sheet (GIS) and the Antarctic Ice-sheet (AIS) have been historically responsible for relatively little sea-level rise to date, but their contribution to sea-level rise has accelerated over the past four decades (currently ~1.20 mm/yr - (Bamber, Westaway, Marzeion, &

Wouters, 2018; Oppenheimer et al., 2019; The IMBIE team, 2018)), increasing by 700% relative to the 1992-2001 period (Bamber et al., 2018; Meredith et al., 2019; The IMBIE team, 2018). With continued warming, however, the potential for significant acceleration of sea-level rise due to ice loss from these two regions poses a large threat to coastal regions globally. While thermal expansion and loss of glaciers outside the GIS and AIS cause irreversible sea-level rise on human timescales due to the ocean's gradual response to warming (Ehlert & Zickfeld, 2018; Solomon, Plattner, Knutti, & Friedlingstein, 2009; Kirsten Zickfeld, Solomon, & Gilford, 2017) and to shifts away from climate conditions that permit mountain glaciers to persist (Lenaerts et al., 2013), their relative contribution to future sea-level change will diminish (Peter U. Clark et al., 2016). Mass losses from the Greenland and Antarctic ice sheets are similarly irreversible and will contribute the majority of expected future sea-level changes.

The GIS and the AIS would drive multi-meter increases in sea-level over future centuries in the event of significant ice-sheet loss (Pattyn et al., 2018). At the same time, an increasing understanding of ice-sheet dynamics has heightened worries that air and sea temperature increases may initiate feedback mechanisms leading to uncontrolled, irreversible ice-sheet collapse over a timescale of centuries (DeConto et al., 2021; J M Gregory, George, & Smith, 2020). Climate change is expected to cause large-scale losses from the GIS and the West Antarctic Ice-sheet (WAIS) at lower levels of climate forcing, followed by further Antarctic ice loss as vulnerable basins of the East Antarctic Ice-sheet (EAIS) retreat under higher levels of warming (Meredith et al., 2019). Given the high concentration of vulnerable human populations and infrastructure in coastal regions worldwide, the risk of crossing critical thresholds for major polar ice-sheets represents a particularly high-priority topic in terms of research, planning, and climate mitigation.

2.3.2 Powerful feedback mechanisms suggest a real possibility for irreversible ice-sheet collapse

Major ice-sheet processes have exerted a dominating influence on sea-levels over the geologic past, driving many meters of sea-level change in response to shifting temperatures (C. P. Cook et al., 2013; Dutton et al., 2015). The GIS, WAIS, and EAIS contain tremendous quantities of ice. Full loss of the GIS would raise global mean sea-level by 7.42 +/- 0.05 m (Morlighem et al., 2017); collapse of the WAIS would yield a global increase of as much as 5.08 m (mean estimate 3.16 m) (L. Pan et al., 2021; Sun et al., 2020). While the EAIS may only begin to lose significant ice mass for higher thresholds of warming, vulnerable marine-terminating basins of the EAIS contain sufficient mass to potentially raise sea-levels globally by up to 19.2 m (Fretwell et al., 2013). However, the stability of portions of the EAIS may depend on the integrity of relatively small volumes of ice at the margins of features such as the Wilkes marine ice-sheet, which could impart tipping behaviour leading to large-scale ice loss over centuries (Mengel & Levermann, 2014).

The magnitude of future sea-level rise resulting from ice-sheet loss depends on climatic and topographic factors as well as differences in glacier characteristics that may trigger ice-sheet destabilization at different critical temperature thresholds for different ice-sheet regions. Generally, ice basins of the GIS and WAIS are more sensitive to current and projected climate change and are hypothesized to likely reach key critical thresholds first, while the EAIS region responds more to higher intensities of warming (N. R. Golledge et al., 2015; Robinson, Calov, & Ganopolski, 2012). Significant ice loss is already occurring for both the GIS and WAIS in the present day, with an ongoing sea-level rise contribution of 1.20 mm/yr (Bamber et al., 2018; Oppenheimer et al., 2019; The IMBIE team, 2018). The EAIS is potentially at a mass balance (no net sea-level contribution) currently thanks to increased snowfall caused by a warming-induced increase in atmospheric moisture, although this balance is subject to considerable temporal variability and uncertainty that do not rule out the possibility of net loss since observations began (Bamber et al., 2018; Boening, Lebsock, Landerer, & Stephens, 2012; Martin-Español, Bamber, & Zammit-Mangion, 2017; The IMBIE team, 2018; Velicogna, Sutterley, & Van Den Broeke, 2014).

The feedbacks affecting the major ice-sheets and the patterns and timeframes of their physical responses to climate change differ markedly between regions. In contrast to the Antarctic sheets, Greenland glaciers typically terminate on land prior to reaching the sea. Consequently, surface melt in response to rising air temperatures that greatly outpaces new ice-sheet accumulation represents the primary cause of ongoing and projected ice loss for the GIS (Bevis et al., 2019). Surface ablation and snow melt can result in the formation of meltwater pools or exposure of darker ice that reduce surface reflectivity, further intensifying melt through a melt-albedo feedback (MacFerrin et al., 2019; Ryan et al., 2019). Refreezing of meltwater also warms adjacent ice. At the same time, meltwater can percolate through holes and crevices through to the foot of the glacier, lubricating the glacier where it meets the bedrock and accelerating the speed of ice flow towards the sea (Bell, 2008). Warming glaciers also exhibit lower viscosity, further increasing their rate of flow. As glaciers move seawards, their altitude falls, bringing them into lower, warmer layers of air - another positive feedback known as the melt-elevation feedback or height-mass-balance feedback (Huybrechts & De Wolde, 1999). Yet while the majority of Greenland ice loss is driven by such atmospheric effects upon surface mass balance (Enderlin et al., 2014; Van Den Broeke et al., 2016), marine-terminating Greenland basins are still subject to ocean-driven forcing (Carroll et al., 2016) and such discharge-driven losses remain significant at the scale of the GIS as a whole (Enderlin et al., 2014). Ocean forcing a focus of ongoing efforts to improve projections of the Greenland ice-sheet's future sea-level contribution (D. A. Slater et al., 2020).

In contrast, marine warming represents the primary driver of Antarctic ice loss, as ice on the AIS margins is typically in direct contact with the ocean. Consequently, such marine-terminating glaciers are at risk of mass loss from processes that result from both oceanic warming (Shepherd, Wingham, & Rignot, 2004) as well as atmospheric warming (Figure 3) (DeConto et al., 2021). The observational record has established the key role of ocean-driven melt in the thinning and retreat of Antarctic ice shelves (Khazendar et al., 2016; Y. Liu et al., 2015; Wouters et al., 2015), although ocean temperatures are themselves subject to atmospheric variability and forcing (Jenkins et al., 2016). Ocean warming in combination with physical stresses can also drive an ice shelf damage feedback in which crevasses and fractures develop within the ice shelves buttressing outlet glaciers of the AIS, accelerating ice loss and further exacerbating damage (Lhermitte et al., 2020).

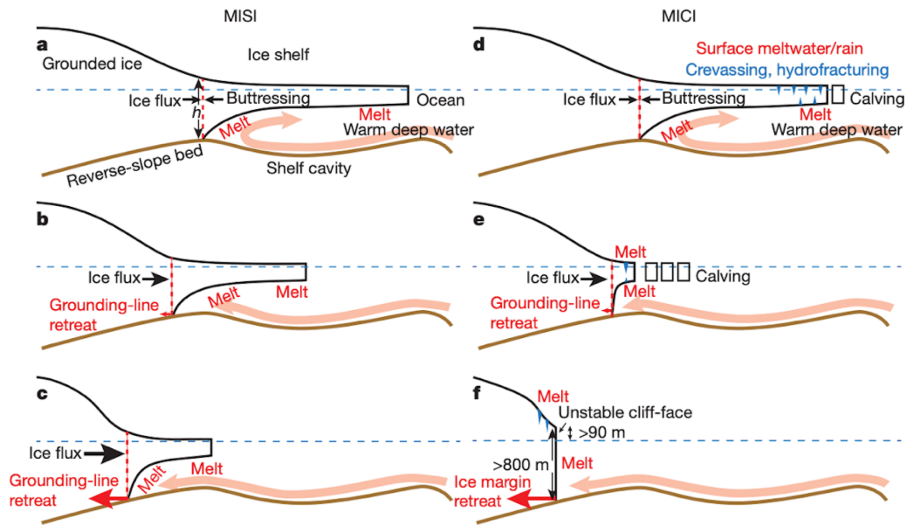

Figure 3: Schematic diagram of the (a, b, c) marine ice shelf instability (MISI) and (d, e, f) marine ice cliff instability (MICI) processes. In MISI, a) ocean and atmospheric heating induce melt proceeding from ice shelf edges, b) causing the ice-sheet's grounding line to retreat onto reverse-sloping bedrock. c) As the height of the glacier above the grounding line grows, the ice-sheet flows out to sea faster in a positive feedback. According to the MICI hypothesis, a) loss of ice shelves due to iceberg calving and surface and subsurface heating leads to b) retreat of the ice-sheet onto reverse-sloping bedrock, followed by c) exposure of progressively higher ice cliff faces that unstably collapse, driving rapid retreat. Figure originally published in (DeConto & Pollard, 2016)**.**

For marine-terminating ice-sheets, the rate at which ice flows out to sea is proportional to the height of the glacier above the grounding line beneath it where the submerged ice-sheet first contacts the bedrock below. On reverse slopes, where the bedrock's height decreases with further distance inland, this relationship results in a positive feedback where the height of the glacier above the grounding line increases as the ice-sheet retreats, which then further accelerates the rate at which the glacier flows out to sea (Schoof, 2007). This irreversible positive feedback mechanism for ice loss is called Marine Ice Shelf Instability (MISI) (Thomas & Bentley, 1978; Weertman, 1974). While not all vulnerable Antarctic glaciers terminate on reverse slopes and are currently thought to be undergoing MISI today, several major basins are currently retreating thanks to processes that may indicate MISI dynamics (Favier et al., 2014; Joughin, Smith, & Medley, 2014; Rignot, Mouginot, Morlighem, Seroussi, & Scheuchl, 2014). In this process, warm subsurface waters cause melt beneath ice shelves, resulting in inland retreat of the grounding line (Shepherd et al., 2004). Should grounding lines retreat beyond forward slopes and onto reverse-sloping topography, as is the potential case for many major Antarctic basins (Ross et al., 2012), the process of MISI would begin rapidly accelerating ice loss from many major glaciers. Much of the WAIS lies on reverse-sloping bedrock well below sea-level (Le Brocq, Payne, & Vieli, 2010), leading to generally higher susceptibility to MISI and greater instability of the WAIS under modest warming scenarios relative to the EAIS (Pattyn et al., 2018).

Warmer waters and atmospheric heating can also cause overlying ice shelves to shrink. In the event that ice shelves disappear completely, a feedback mechanism known as Marine Ice Cliff Instability (MICI) may trigger at locations where the height of cliffs at an exposed ice-sheet's edge exceeds critical thresholds. Beyond such heights, the shear strength of ice would be insufficient to withstand longitudinal stress at the cliff face, causing progressive collapse via iceberg calving (Bassis & Jacobs, 2013; Bassis & Walker, 2012). Since the thickness of ice shelves often increases further inland, MICI could represent a strong positive feedback for ice-sheet loss. Furthermore, unlike MISI, MICI may be capable of occurring on flat or forward slopes so long as the ice cliff threshold is exceeded (Pollard, DeConto, & Alley, 2015). As ice shelves also provide a supportive "buttressing" effect that opposes and slows the rate of ice flux to sea, loss of ice shelf mass itself also helps speed the glacier's progress towards the ocean (Schoof, 2007). A recent modelling analysis suggests that over realistic timescales of ice shelf collapse, physical stress and deformation are more evenly distributed across the cliff face, potentially preventing widespread cliff failure as proposed by the MICI hypothesis (Clerc, Minchew, & Behn, 2019). However, this does assume that the ice shelf is not otherwise damaged or compromised, in which case thresholds for stable cliff heights may decrease significantly (DeConto et al., 2021). Currently, no evidence indicates that MICI is occurring on Antarctic ice-sheets, and deep uncertainty remains concerning the potential importance of MICI for projections of future ice loss (Meredith et al., 2019). Overall, projecting ice-sheet losses from MICI remains a contentious topic given its significant impact on near-term and long-term sea-level rise estimates. While studies incorporating MICI dynamics perform well in reproducing past ice sheet retreat under warm paleoclimates (DeConto et al., 2021), researchers have not yet fully validated the physical mechanisms involved using field observations (Edwards et al., 2019).

Surface mass loss via atmospheric heating represents a major threat to Antarctic glaciers as well. Atmospheric warming may threaten smaller ice shelves and tongues that currently help stabilize glaciers (Schoof, 2007), thus increasing the threat MISI and potentially MICI pose to predominantly marine-terminating Antarctic ice-sheets over the coming centuries. Some recent modeling work has also suggested that the majority of future sea-level rise resulting from Antarctic ice mass loss could occur due to atmospheric warming rather than ocean heating, due to amplification of MISI/MICI dynamics in addition to surface melt and associated meltwater feedbacks (DeConto et al., 2021; Sun et al., 2020).

At the same time, a handful of negative feedbacks may somewhat slow rates of Antarctic ice-sheet loss. For instance, increasing ice congestion in restricted bays may provide added resistance, buttressing some glaciers and inhibiting ice movement out to sea (Seneca Lindsey & Dupont, 2012), although this does not protect basins that terminate in the open sea, such as the Thwaites (DeConto et al., 2021). Increased snowfall over portions of Antarctic ice have also provided mass inputs that offset a portion of ice loss (Boening et al., 2012). These factors only represent relatively minor influences upon the ice-sheets' mass balance as a whole. More significantly, uplift of the land surface beneath retreating ice-sheets as the weight of the overlying ice is reduced has been suggested to act much faster than previously assumed, potentially providing a negative feedback (Barletta et al., 2018), although uplift may also accelerate meltwater flux into the ocean and further amplify global sea level rise over coming centuries (L. Pan et al., 2021). Overall, however, modeling suggests that negative feedbacks will exert limited impact in slowing Antarctic ice sheet retreat (DeConto et al., 2021).

Overall, the balance of current trends acts to drive the GIS and WAIS increasingly towards net feedbacks leading to sustained, irreversible ice-sheet loss. Evidence from geological observations, paleoclimate records, surveys of glaciers and bed topography, and modeling analyses strongly support the conclusion that warming beyond key thresholds will initiate ice-sheet collapse over a time scale of centuries. Sea-level rise in response to loss of the Greenland ice-sheet and major portions of the Antarctic ice-sheet therefore represents an extreme and credible threat to coastal regions worldwide.

2.3.3 Commitment to loss of the Greenland and West Antarctic ice-sheets yields accelerated, multi-meter sea-level rise over centuries and millennia

Ongoing field and satellite studies have confirmed or even revised upward earlier findings that vulnerable ice-sheets are already feeling the effects of climate change (Hamlington et al., 2020). Mass loss from Greenland has continued, with the pace of its contribution to sea-level having accelerated over time in recent decades (McMillan et al., 2016). The current total contribution of ice-sheet loss to sea-level rise is following that predicted by upper-end AR5 emissions scenarios, although this results from a combination of observed surface mass balance losses that have exceeded the range of modelled losses and observed losses via ice dynamics that have fallen largely in the mid-to-lower range of model projections (T. Slater, Hogg, & Mottram, 2020). Individual ice shelves on the West Antarctic Peninsula, the Southern Antarctic Peninsula, and in the Bellingshausen and Amundsen seas are experiencing losses via behavior potentially consistent with irreversible onset of MISI (Paolo, Fricker, & Padman, 2015; Rignot et al., 2014; Wouters et al., 2015). Recent observational evidence also suggests that the EAIS may be worryingly more vulnerable to ocean heating than assumed previously (Silvano, Rintoul, & Herraiz-Borreguero, 2016). New research has also pointed to a potentially longer-term natural trend of ice-sheet thinning for parts of the EAIS over the past 300 years, potentially increasing the region's vulnerability to future ice loss (W. A. Dickens et al., 2019).

Initiation of irreversible ice-sheet collapse would bring about serious long-term consequences for global sea-level. For summer warming of more than 2°C, the GIS has been assessed as more likely to reach a critical threshold than not (Pattyn et al., 2018). Local annual mean warming of 3°C could be attained this century and would eliminate nearly all Greenland glaciers, raising seas by 7 m over approximately 1,000 years (Jonathan M. Gregory, Huybrechts, & Raper, 2004; Huybrechts, Letreguilly, & Reeh, 1991). However, a new study coupling an ice-sheet model with an atmospheric general circulation model has produced results suggesting that the Greenland Ice Sheet may not exhibit bistable behavior in response to transgression of a critical threshold (J M Gregory et al., 2020). While further research is required to verify the existence of intermediate diminished states of the Greenland Ice Sheet under steady-state warmer climates, this result leaves open a possibility that the Greenland Ice Sheet may exhibit stable states between its modern extent and complete loss.

For the Antarctic ice sheet, a modeling analysis utilizing the hysteresis approach (S. Rahmstorf & England, 1997) observed strong retreat of grounding lines in the WAIS for global temperatures 1-2.5°C above pre-industrial, producing a partial collapse of the WAIS at a critical threshold around 2°C warming (Garbe, Albrecht, Levermann, Donges, & Winkelmann, 2020). Under a worst-case RCP8.5 emissions scenario incorporating a strong but uncertain MICI feedbacks (DeConto & Pollard, 2016), modelers have calculated a potential for Antarctic ice loss to contribute 34 cm (17th-83rd percentiles: 20 to 53 cm) of sea-level rise by 2100, with an eventual sea-level increase of 9.57 m (6.87 to 13.55 m) by 2300 (DeConto et al., 2021). 3C pathways result in significant WAIS ice loss within 500 years, driven strongly by intense retreat of the Thwaites Glacier, yielding 15 cm (8 to 27 cm) of sea-level rise from Antarctica by 2100 and an eventual total Antarctic contribution of 1.54 m (1.04 to 2.03 m) by 2300 (DeConto et al., 2021).

Modeling results not incorporating MICI dynamics generally produce somewhat more conservative sea-level changes, with an Antarctic contribution to sea-level rise of 22 cm of sea-level rise by 2100 (5th-95th percentiles: 8 to 43 cm) estimated using pessimistic assumptions under the high-emissions RCP8.5 pathway (Edwards et al., 2021). Newer ISMIP6 ice-sheet model simulations using CMIP5 model outputs have thus far yielded a similar range of -6.7 to 35 cm of sea-level rise contributed by the Antarctic Ice Sheet by 2100 under RCP8.5 (Seroussi et al., 2020). Under the high-emissions RCP8.5 forcing, one modeling study assessed a cumulative contribution from Greenland and Antarctica of 25 cm of sea-level rise by 2100 (Nicholas R. Golledge et al., 2019). Overall, total predicted sea-level rise by 2100 generally ranges between 0.61 and 1.10 m for RCP8.5 (Oppenheimer et al., 2019).

Projections of future sea-level rise from ice-sheet losses remain highly uncertain, primarily due to limited observational records, incomplete understanding of ice-sheet dynamics, and model limitations. Changes to buttressing from ice shelves and the dependence of mass losses upon sliding and basal friction physics will strongly determine sea level contributions from Antarctica (Sun et al., 2020). Estimates of ice-sheets' contribution to sea-level by 2090 evident across a selection of published literature span a fourfold range (A. M. R. Bakker, Louchard, & Keller, 2017), with structured expert judgement yielding a similarly wide range of estimates (median estimate: 0.51 m, 95th percentile estimate: 1.78 m) (Bamber, Oppenheimer, Kopp, Aspinall, & Cooke, 2019). Model projections thus produce variable results for the magnitude and distribution of future regional sea-level rise if different assumptions are used to model Greenland and Antarctic ice loss (Kopp et al., 2017), highlighting the relevance of current knowledge gaps to policymakers. Such uncertainty poses a significant challenge to governments and planners' attempts to assess adaptation needs (Rasmussen, Buchanan, Kopp, & Oppenheimer, 2020).

Antarctic ice-sheet collapse would be irreversible absent substantial cooling that would permit reformation of buttressing ice shelves. Otherwise, retreat would continue until topography that halts MISI/MICI is reached. Greenland melt is theoretically reversible to a point, with the ice sheet regrowing to its present-day extent upon a modeled restoration of late 20th-century climate provided ice loss is confined to a magnitude of 4.0 m sea-level equivalent or less (J M Gregory et al., 2020). Beyond this threshold, the retreat of ice from northern Greenland reduces regional snowfall, preventing regrowth of ice around 2.0 m of sea-level equivalent in total. The albedo of the ice sheet's surface remains a relatively uncertain factor in modeling the surface mass balance of the GIS, with higher recovery of the ice sheet possible if surface albedo remains high but improbable at lower albedo values.

Timescales of collapse depend strongly on bed topography of individual ice-sheet basins and can require centuries to millennia. The Thwaites glacier, for instance, may already have reached the early stages of irreversible mass loss, although its rate of reduction will likely remain moderate over the 21st century, with collapse potentially occurring over a period of 200-500 years (Cornford et al., 2015; DeConto et al., 2021; N. R. Golledge et al., 2015). EAIS basins may take more than a century under warming scenarios to begin shedding significant mass, with multi-meter contributions to sea-level rise over the course of several millennia (N. R. Golledge et al., 2015). As mentioned above, reduction of the GIS will likely require a millennium. Consequently, under our current best understanding, ice-sheet collapse cannot generally be considered an abrupt or fast phenomenon, although sustained high rates of sea-level rise (>1 cm/yr by 2200, with further acceleration to up to a couple centimeters per year beyond) may seriously strain coastal adaptation efforts (Oppenheimer et al., 2019).

At the same time, models indicate that strong climate mitigation may avert significant fractions of potential sea-level rise and prevent ice-sheet collapse across large regions. Under RCP2.6, a modeling experiment incorporating an aggressive MICI mechanism produced virtually no net Antarctic contribution to global mean sea-level by 2100, with only a 20 cm rise by 2500 (DeConto & Pollard, 2016). Another study found that the RCP2.6 scenario prevented collapse of the WAIS regardless of uncertainties in model parameters (Bulthuis, Arnst, Sun, & Pattyn, 2019). The IPCC's synthesis of scientific findings in the SROCC suggests total sea-level rise by 2100 under RCP2.6 could be contained to a range of just 0.29 to 0.59 m (Oppenheimer et al., 2019). More recent modeling efforts indicate that limiting global mean warming to 1.5°C might reduce the Antarctic contribution to global sea level rise by 2100 to 13 cm (Edwards et al., 2021). Although significant uncertainties remain regarding the precise temperature thresholds that could trigger ice-sheet collapse, research to date strongly suggests that aggressive climate mitigation in accordance with Paris Agreement goals significantly limits risks from ice-sheet instabilities.

2.4 Carbon release from thawing permafrost

2.4.1 Background

Permafrost is any earth material (soil, sediment, rock) that remains below freezing temperatures for at least several consecutive years, with many permafrost regions worldwide having remained predominantly frozen over the past several thousand years. Due to minimal rates of decomposition at such low temperatures, considerable quantities of organic matter have accumulated and become incorporated in these frozen soils. Consequently, sizable stocks of carbon lie contained within permafrost soils globally. Today permafrost covers ~23 million km^2 of the planet, with 13–18×106 km^2 in the Arctic, 1.06×10^6 km^2 in the Tibetan plateau and 16–21×106 km^2 in subsea and Antarctic regions (Chadburn et al., 2017; Gruber, 2012; Sayedi et al., 2020; D. Zou et al., 2017). Total organic carbon content of all permafrost soils in the Northern Hemisphere are assessed to range between 1460 - 1600 Gt C, nearly twice the amount of carbon currently in the atmosphere (E. A. G. Schuur, McGuire, Romanovsky, Schädel, & Mack, 2018). On a worldwide scale, permafrost carbon represents about one-third of all global soil carbon within the upper 3m (Jobbágy & Jackson, 2000; E. A.G. Schuur et al., 2015).

Rising global temperatures are causing permafrost to warm (Biskaborn et al., 2019; Lewkowicz & Way, 2019). Temperatures at permafrost depths 10-20 m below the ground surface are now 2-3°C higher than those observed 30 years before (Figure 4) (V. Romanovsky et al., 2017). Permafrost within the colder continuous permafrost zones of the far north warmed by 0.29 +/- 0.15°C between 2007 and 2016, highlighting that permafrost with mean annual temperatures well below freezing is still vulnerable to thaw due to factors such as landscape morphology and ground ice content (Biskaborn et al., 2019). Over the past two decades, the climate community has devoted serious efforts towards assessing how widespread permafrost thaw will stimulate organic matter mineralization and C emissions, a potential positive feedback to climate termed the permafrost carbon feedback to climate (PCF). This feedback depends on understanding how much of the permafrost carbon pool is vulnerable to mineralization, how quickly permafrost carbon losses will occur over time, whether these emissions will be released as CO_2 versus CH_4, and the degree to which permafrost carbon loss can be offset by increasing plant biomass post-thaw.

In addition to warming, changes in weather patterns such as elevated precipitation rates also influence thermokarst processes (Kokelj et al., 2015). Observations indicate that high-latitude rainfall may already be increasing (J. E. Walsh, Overland, Groisman, & Rudolf, 2011), and climate models predict a strong likelihood for future high-latitude precipitation to increase with climate change (Overland et al., 2011). Higher temperatures in permafrost regions could also boost the frequency of wildfires, both releasing significant carbon and transferring heat to deeper permafrost layers (Goetz, MacK, Gurney, Randerson, & Houghton, 2007; Holloway et al., 2020; Mack et al., 2011; Randerson et al., 2006; X. J. Walker et al., 2019). Strong evidence points towards an increasing frequency and severity of wildfires throughout the arctic and boreal north

(Flannigan, Stocks, Turetsky, & Wotton, 2009; Hanes et al., 2019; Kasischke & Turetsky, 2006; McCarty, Smith, & Turetsky, 2020). Field observations have demonstrated that wildfire can act as a major driver of regional permafrost thaw, with fire contributing towards the expansion of thermokarst bog area in western Canada over decades (Gibson et al., 2018).

Permafrost thaw - and resulting carbon emissions - has sometimes been characterized as a fast-acting tipping element with tipping points which, once crossed, would trigger abrupt and severe warming. On the other hand, permafrost is still perceived as a relatively stable component of the climate system on century or possibly millennial times scales. Thus, whether permafrost carbon will or will not be an important factor in abrupt carbon cycle shifts in addition to the other components reviewed here is contentious (Dyonisius et al., 2020; Petrenko et al., 2017). Recent modeling studies suggest that permafrost carbon release will take effect after 2100 and will be greatly reduced with more stringent climate mitigation policy (McGuire et al., 2018; M. Turetsky et al., 2020). The majority of permafrost thaw will occur via active layer thickening, often referred to as gradual permafrost thaw because it affects centimeters of surface permafrost relatively slowly on a time scale of decades to centuries (McGuire et al., 2018; Schneider Von Deimling et al., 2015). Abrupt permafrost thaw processes (thermokarst) can affect meters of permafrost over more rapid timescales locally (days to years), but like gradual thaw will cumulatively contribute towards climate change over a century or longer rather than in a single event (M. Turetsky et al., 2020; M. R. Turetsky et al., 2019). This is because while abrupt thaw represents an important ecosystem state change (M. G. Turner et al., 2020), the geophysical and ecological conditions required for thermokarst limit abrupt thaw processes to less than 20% of the Arctic region (Olefeldt et al., 2016; M. Turetsky et al., 2020). Even beyond 2100, methane fluxes from thermokarst will likely induce only a modest to moderate climate forcing (29-78 Mt C/yr through 2300 under a worst-case RCP8.5 scenario (M. Turetsky et al., 2020)) relative to anthropogenic methane emissions today (335 Mt C/yr).

Nevertheless, this added input of atmospheric carbon (worst-case estimates equivalent to up to ~400 Gt C by 2300) (M. R. Turetsky et al., 2019) would be climatically significant, further incentivizing climate change mitigation to a level that would minimize the impact from this feedback. Remaining uncertainty concerning feedbacks between temperature, ground ice, precipitation, permafrost thaw, local landscape and vegetation changes, and the response of microbial communities and biomass (Bonnaventure et al., 2018; Jorgenson et al., 2010; Kokelj et al., 2015; McGuire et al., 2018; Rivkina et al., 2004; Shur & Jorgenson, 2007) also leaves considerable error associated with net CO_2 and methane emissions estimates (Dean et al., 2018).

Beyond carbon and climate feedbacks, permafrost thaw also has important consequences for local communities, infrastructure, and ecosystems ranging from structural damage and loss of life caused by unstable ground to an elevated risk of infectious disease outbreaks at the hands of live pathogens liberated from thawed permafrost (Meredith et al., 2019; M. G. Walsh, De Smalen, & Mor, 2018). For example, abrupt thaw was associated with the majority of hazards experienced by land users in Alaska, impacting the way northerners can hunt or travel over land or rivers (Gibson, Brinkman, Cold, Brown,

& Turetsky, 2021). Models suggest that ambitious climate mitigation efforts would appreciably reduce the extent of global permafrost loss (Chadburn et al., 2017; McGuire et al., 2018). Consequently, limiting the extent of climate change represents a powerful approach for reducing both the climate risks associated with permafrost thaw as well as impacts upon northern communities.

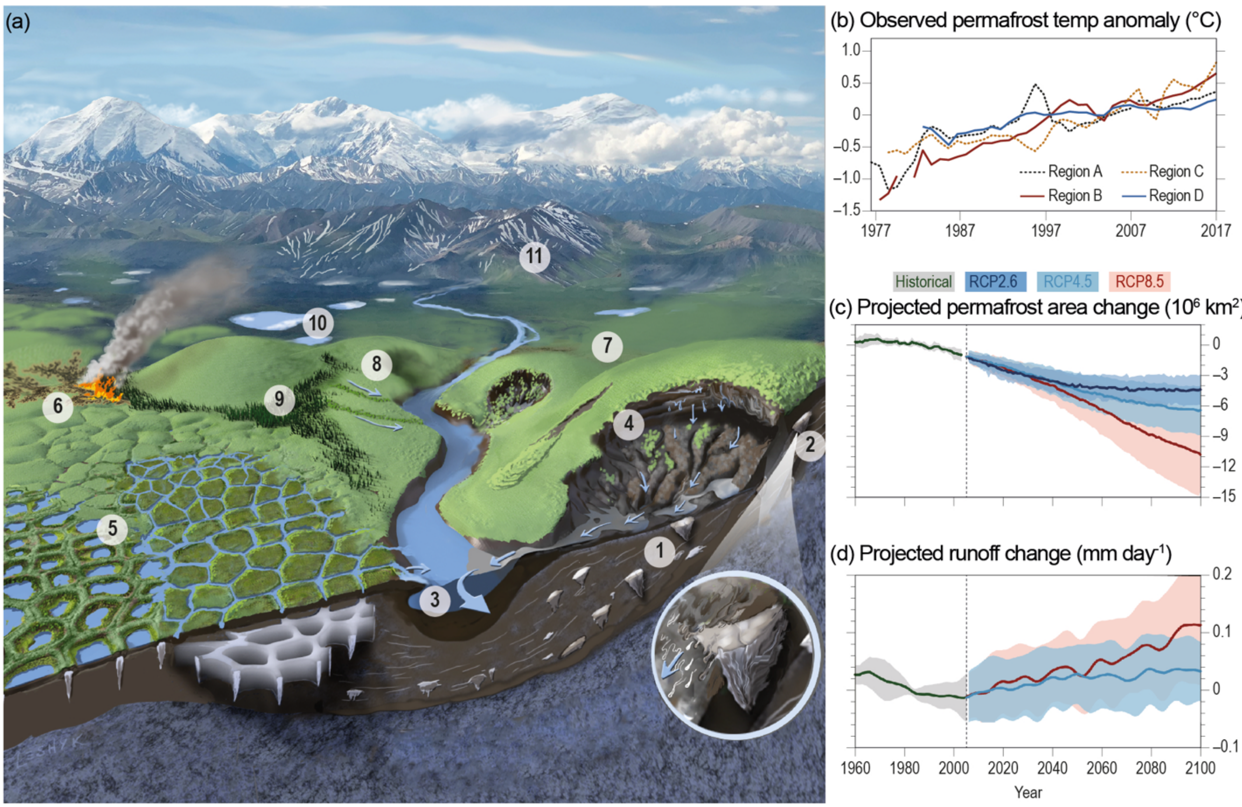

Figure 4: a) Diagram illustrating components of the circumpolar Arctic terrestrial landscape potentially sensitive to climate change, including 1) permafrost, 2) ground ice, 3) river flow rates, 4) abrupt thaw via landslides and small-scale processes, 5) surface water accumulation, 6) wildfires, 7) tundra biome shifts, 8) shrub vegetation, 9) boreal forests, 10) lake ice, and 11) seasonal snow cover. b) Observed permafrost temperature anomalies (relative to the 2007-2009 International Polar Year baseline) over time (V. E. Romanovsky et al., 2017)**. c) global area of near-surface permafrost soils and d) projected change in surface runoff to the Arctic Ocean under low (RCP2.6), middle-of-the-road (RCP4.5), and worst-case (RCP8.5) emissions scenarios. Figures adapted with permission from Figure 3.10 in the IPCC SROCC Ch 3** (Meredith et al., 2019)

2.4.2 Cumulative methane and CO_2 release from abrupt and gradual permafrost thaw manifests over a century or more

The potential for permafrost change to govern C fluxes between the biosphere and the atmosphere depends upon the response of gradual and abrupt permafrost thaw processes to rising temperatures, changing weather (including precipitation patterns), and shifts in patterns of disturbance, followed by impacts on C cycle processes including the microbial decomposition of organic matter and either the production of carbon dioxide under a range of moisture and redox conditions or the production of methane under anoxic saturated conditions. Patterns of vegetation and biomass recovery patterns post-thaw also are important to ecosystem C balance including the potential for new peat accumulation post-thaw. Summed up over the entirety of northern permafrost regions, the potential of such factors to significantly influence overall methane release remains relatively poorly constrained.

It is highly likely that a sizable fraction of permafrost area globally will thaw by end-of-century, though estimates of the extent of loss range widely from 5% to 70% (C D Koven et al., 2015; D. M. Lawrence, Slater, & Swenson, 2012; McGuire et al., 2018; Schaefer, Zhang, Bruhwiler, & Barrett, 2011; E. A.G. Schuur et al., 2015; Edward A.G. Schuur & Abbott, 2011; Wisser, Marchenko, Talbot, Treat, & Frolking, 2011). Some permafrost can persist even in the face of disturbance and 1.5-2°C of observed regional warming (Holloway & Lewkowicz, 2020; James, Lewkowicz, Smith, & Miceli, 2013). For example lowland permafrost associated with thick peat layers persisted as relict patches of permafrost remnant from the Little Ice Age in North America (Halsey, Vitt, & Zoltai, 1995). In large part due to ecosystem protection of permafrost created by vegetation, peat, and snow characteristics, thermal modelling also suggests that permafrost in the southern margins of the permafrost zone may be more resilient than previously assumed (Way, Lewkowicz, & Zhang, 2018). However, such findings also emphasize that substantial permafrost degradation at regional to continental scales has already occurred to date and will continue under high and mid-range emissions scenarios.

In addition to C emissions associated with widespread gradual thaw, processes driving abrupt permafrost thaw are expected to contribute meaningfully to global climate. Abrupt thaw could produce CH_4 and CO_2 fluxes (38-104 Mt CH_4/yr, 0.01-0.45 Gt C/yr of CO_2 through 2300 under an RCP8.5 scenario, see Section 3.5) with a comparable radiative forcing impact to that produced by C emissions from gradual thaw expected later this century (McGuire et al., 2018; Schneider Von Deimling et al., 2015; Schneider von Deimling et al., 2012; M. Turetsky et al., 2020)). Abrupt thaw processes - the collective term for thermokarst (land slumping and subsidence), rapid erosion, and similar phenomena - lead to more abrupt exposure and thaw of permafrost on time scales of days to years (Benjamin W. Abbott & Jones, 2015). Thermokarst-prone landscapes can occur in both discontinuous and continuous permafrost (Fraser et al., 2018; Lewkowicz & Way, 2019; Olefeldt et al., 2016; Rudy, Lamoureux, Kokelj, Smith, & England, 2017) and are often associated with surface water accumulation and/or melting ground ice that triggers sudden landslides, crevasses, or subsidence. Carbon mobilized by thermokarst events in particularly old Yedoma permafrost soils (>21 000 years old) has demonstrated rapid rates of biodegradation, underlining the potential for significant carbon release from thermokarst features (Vonk et al., 2013). On the other hand, other types of permafrost carbon, such as Holocene aged permafrost soils in some peatlands, have been found to be relatively resistant to degradation post-thaw (Heffernan, Estop-Aragonés, Knorr, Talbot, & Olefeldt, 2020).

Currently, permafrost carbon emissions are dominated by high CO_2:CH_4 ratios, though methane fluxes from abrupt thaw have received significant media attention. Analysis of historic methane concentrations during periods of past warming have suggested that release of methane from permafrost and marine gas hydrates during the last deglaciation was likely small (<19 Mt CH_4/yr), although the present-day Arctic is warming at a much faster rate than that previously experienced in the region (Dyonisius et al., 2020; Petrenko et al., 2017). Empirical and modeling studies agree that permafrost C emissions currently are dominated by CO_2 fluxes in oxic soil environments, with much smaller fluxes of CH_4 occuring in anoxic conditions (Schädel et al., 2016; Walter Anthony et al., 2018). Particularly in lowland regions prone to abrupt thaw, the contribution of CH_4 to total permafrost C release is expected to increase in the future (M. Turetsky et al., 2020; Walter Anthony et al., 2018). Increasing methane emissions from saturated thawing soils in these northern lowlands often is driven by modern carbon assimilated recently by the ecosystem rather than older carbon that previously resided in permafrost layers (Estop-Aragonés et al., 2020) highlighting the important connection between permafrost thaw, land cover change, and ecosystem carbon emissions. The role of methane oxidation has been shown to be important in some thaw environments (B. W. Abbott, Jones, Godsey, Larouche, & Bowden, 2015) but requires more study to determine its overall significance to permafrost carbon emissions as well as its sensitivity to future changes in permafrost, climate, and biophysical conditions. Beyond microbial transformations of methane, natural methane stores trapped under currently-frozen soil layers may also be released thanks to permafrost thaw, although this source is not currently anticipated to be climatically significant (Petrenko et al., 2017). Emissions of nitrous oxide - another potent greenhouse gas - from permafrost also may be non-negligible (Voigt et al., 2020; Wilkerson et al., 2019) and require further study. In general, improved projections of hydrological changes within the permafrost region (Andresen et al., 2020) and better quantification of the rates of permafrost organic carbon mineralization into CO_2 versus CH_4 (or other greenhouse gases such as N_2O), and the fate of permafrost C exported as dissolved organic matter in aquatic environments remain active areas of study with major climate implications (Bowen, Ward, Kling, & Cory, 2020; Laurion, Massicotte, Mazoyer, Negandhi, & Mladenov, 2020; Zolkos & Tank, 2020).

At least some portion of permafrost C release is likely to be offset by increased plant growth and new accumulation of biomass (Charles D. Koven, Lawrence, & Riley, 2015; McGuire et al., 2018; M. Turetsky et al., 2020; M. R. Turetsky et al., 2019). Landscape to regional vegetation shifts are occurring as a result of permafrost thaw (Baltzer, Veness, Chasmer, Sniderhan, & Quinton, 2014) but may in turn also carry important consequences for permafrost carbon cycle feedbacks. In particular, the large-scale expansion of shrubs across the circumpolar north (Bjorkman et al., 2018; Tape, Sturm, & Racine, 2006) may accelerate the seasonal timing of snow melt, acting as a positive feedback upon permafrost thaw (Wilcox et al., 2019). Plant roots can further stimulate the decomposition of organic carbon in thawed permafrost, exerting a potentially large impact on regional scales as warming and enhanced nitrogen availability promotes vegetation growth (Albano, Turetsky, Mack, & Kane, 2021; Keuper et al., 2020). Simulations focused on permafrost thaw show that enhanced plant carbon uptake can entirely offset permafrost carbon emissions under the RCP4.5 emissions scenario (McGuire et al., 2018). The future of Arctic vegetation and CO_2 uptake depends on patterns of regional wetting versus drying in combination with disturbances like fire regimes. It seems likely that ecosystem carbon balance in the northern permafrost region will be strongly affected by the strength of plant CO_2 fertilization in combination with vegetation response to climate, which need better mechanistic representation in most large scale models. The response of different landscapes and vegetation types to warming, changes to nutrients, increased CO_2, precipitation, and fire will generally drive large-scale ecosystem changes across the circumpolar north, with important implications for not only permafrost but also northern boreal forests (see Section 2.5).

Projections for emissions from thawed permafrost by 2100 are quantitatively large (range of 37-174 Gt C, with two recent estimates of 57 Gt C and 87 Gt C under a worst-case RCP8.5 emissions scenario) (C D Koven et al., 2015; Meredith et al., 2019; Schneider Von Deimling et al., 2015). These greenhouse gas fluxes do possess climate significance. A multi-model mean of 92 Gt C released from gradual thaw by 2100 under RCP 8.5 (Meredith et al., 2019) would represent about a decade of present-day emissions (~10 Gt C/yr) (Peters et al., 2019). Most carbon release from gradual permafrost thaw occurs on time scales of several centuries, with more substantial net carbon emissions occurring after 2100 (McGuire et al., 2018). Over longer multi-century timescales, models demonstrate higher uncertainty, with estimates for carbon release from gradual thaw by 2300 under RCP8.5 ranging between 167 Gt C absorbed to 641 Gt C released (mean of 208 Gt C released) (McGuire et al., 2018). Thermokarst processes could emit an additional 60-100 Gt C by 2300 under RCP8.5, although associated climate impacts could vary depending on the proportion of methane emissions (M. Turetsky et al., 2020; M. R. Turetsky et al., 2019). Climate mitigation could significantly preserve existing permafrost carbon, with the modest-mitigation RCP4.5 scenario significantly minimizing permafrost carbon release from both abrupt and gradual thaw (McGuire et al., 2018; M. Turetsky et al., 2020). However, the near-term (21st century) versus longer-term (through 2300) sensitivity of the CO_2 and CH_4 feedbacks to temperature may differ in an RCP4.5 versus an RCP8.5 scenario, with a potential for nonlinear increases in sensitivity over time (M. Turetsky et al., 2020).

Efforts to assess the climate implications of future permafrost thaw are complicated by the limitations of current models (Bonnaventure et al., 2018). CMIP5 models produce a wide range of estimates due to differences in modelled transfer of warming to permafrost and to varying model treatment of surface factors such as snow and hydrology (Charles D. Koven, Riley, & Stern, 2013). Models' limited ability to account for ground ice represents a particularly important challenge, as ground ice may slow rates of permafrost warming over coming decades (H. Lee, Swenson, Slater, & Lawrence, 2014) but ultimately increase landscape susceptibility to thermokarst over longer timescales (Nitzbon et al., 2020). CMIP6 models generally yield similar projections for future changes in permafrost area and volume, but still do not incorporate important factors including ground ice (E. J. Burke, Zhang, & Krinner, 2020). Land surface models continue to exhibit wide spread in estimates of soil carbon stocks and in the response of soil carbon dynamics to changing temperature and moisture (Huntzinger et al., 2020). The use of high-resolution regional permafrost models forced by the results of GCM runs may enable improvements in model performance, by better capturing landscape heterogeneity (Lara et al., 2020).

In summary, permafrost carbon release might be more accurately characterized as a slow rather than an abrupt tipping element. Overall, carbon release from permafrost takes place over timescales of a century to several centuries. Research to date has not identified any permafrost feedbacks with sufficient scale to drive methane emissions that could substantially influence global climate within an abrupt time scale of years (Dean et al., 2018; National Research Council, 2013). Yet even if the permafrost carbon cycle feedback does not represent an extreme runaway warming threat, it nevertheless carries appreciable importance in the context of climate change. Permafrost feedbacks do pose a serious challenge to ambitious mitigation targets but are unlikely to independently warm global climate by a degree or more within this century (also see Section 3.4).

2.5 Large-scale boreal forest ecosystem shifts

2.5.1 Background

The northern boreal biome has felt the effects of climate change to a substantial degree, with decades of observations highlighting rapid regional temperature increases, more frequent occurrence of wildfires, and intensification of pest-driven tree mortality. Near-surface fall and winter air temperatures within the band of latitudes from 70-90°N rose by 1.6°C per decade over the 1989-2008 period, with summer temperatures rising at a rate of 0.5°C per decade (Screen & Simmonds, 2010). The observational record has seen a marked increase in the extent, frequency, and severity of boreal wildfires, with this increasing trend predicted to continue with further warming (De Groot, Flannigan, & Cantin, 2013; Flannigan et al., 2009; Seidl et al., 2020; Veraverbeke et al., 2017). Warmer temperatures have additionally been linked to acute outbreaks of insects leading to large-scale tree mortality events in Alaska, Canada, and Siberia (Boyd et al., 2021; Kharuk, Im, & Soldatov, 2020; Kurz, Stinson, Rampley, Dymond, & Neilson, 2008; Sherriff, Berg, & Miller, 2011; US Forest Service, 2019), sparking concern that similar pest invasions could occur more often in the future, and into novel hosts and ranges (de la Giroday, Carroll, & Aukema, 2012).

Between the biomass in soil, permafrost, and living and dead vegetation, boreal forests represent a significant pool of terrestrial organic carbon (30% of global soil carbon) (Mcguire et al., 2009; M. R. Turetsky et al., 2019), and constitute 30% of global forest area (Kasischke, 2000). Of this fraction, two-thirds of boreal forest are found within Russia, with Russia's boreal forests estimated to contribute around half (0.6 Gt C/yr) of the total global terrestrial carbon sink (Dolman et al., 2012; Schaphoff et al., 2013). Recent research has proposed that boreal forest carbon stocks could be underestimated, with updated calculations suggesting that boreal regions hold more terrestrial carbon (Bradshaw & Warkentin, 2015) than tropical areas, which have been previously suggested to harbor the most carbon globally among all biomes (Y. Pan et al., 2011).

It is important to note that 95% of boreal zone carbon is stored in peatlands and soils (Bradshaw & Warkentin, 2015), and an uncertain proportion of this underground carbon stock overlaps with the stock of permafrost carbon (Section 2.4). Boreal vegetation and permafrost soils also interact with one another (Carpino, Berg, Quinton, & Adams, 2018), complicating attempts to disentangle potential carbon fluxes associated with changes in the vegetation and soil systems. Regardless, boreal forests make up an important component of the terrestrial and global carbon cycles, with changes in this biome potentially acting as large climate feedbacks.

A growing number of research studies over the past 25 years have provided increasing evidence for the possibility of region- (e.g. North America versus Siberia) and landscape- (e.g. upland vs. lowland) specific shifts within the circumpolar boreal biome, with potentially significant climate implications (Chapin et al., 2005; Foley, Kutzbach, Coe, & Levis, 1994; Lenton et al., 2008; X. Lin et al., 2020; Scheffer et al., 2012). This hypothesis theorizes that a landscape-dependent mix of temperature, moisture, and precipitation changes, shifts in wildfire regimes and soil conditions, greater vulnerability to pest insect outbreaks, and a lengthening of the growing season may all combine to drive acute mortality of boreal forest vegetation in drier, upland areas and at the southern margins of boreal zones and its replacement by more open deciduous woodlands and grasslands, as reviewed by (Gauthier, Bernier, Kuuluvainen, Shvidenko, & Schepaschenko, 2015) and predicted in forest modeling studies (Foster et al., 2019; Mekonnen, Riley, Randerson, Grant, & Rogers, 2019). In contrast, increasing boreal forest productivity in wetter ecosystems and at latitudinal and elevational range boundaries is also possible (Foster et al., 2019; Pastick et al., 2019). Region-specific differences in wildfire regimes and permafrost dynamics (Rogers, Soja, Goulden, & Randerson, 2015) may also lead to diverging responses to climate change. (X. Lin et al., 2020) found that the net sink of atmospheric carbon in Siberia has increased by 18 Tg/year since 1980, while the carbon flux over the North American boreal and arctic biomes has remained relatively constant. Such ecosystem- and continental-scale shifts could represent a climatically significant tipping element with the potential to change land surface albedo (Beck et al., 2011; Bonan, Pollard, & Thompson, 1992) over large areas and release a considerable pool of carbon to the atmosphere, in addition to altering the strength of an important terrestrial organic carbon sink.

Considerable uncertainties regarding the resilience of boreal forests to climate change and the mechanisms driving potential ecosystem shifts remain, making it difficult to determine the likelihood that this system could act as a tipping element (Schaphoff, Reyer, Schepaschenko, Gerten, & Shvidenko, 2016). The observed and predicted reductions in boreal forest area in southern and upland boreal regions combined with the predicted northward treeline expansion in response to warmer temperatures leads to multiple competing, complex climate impacts (Beck et al., 2011; Foster et al., 2019; Ju & Masek, 2016; Pastick et al., 2019; Pearson et al., 2013). Nevertheless, the vast size of the boreal biome, the near-certainty that the region will undergo imminent changes this century that are large relative to other parts of the world, and the strong observational evidence for intensification of disturbances to the boreal forest suggest that future trends in boreal ecosystems could play an important role in determining the trajectory of global climate.

2.5.2 Boreal forests worldwide are experiencing rapid changes with the potential for northward expansion and southern margin shifts to deciduous forest or grasslands

Observations of ongoing changes in boreal ecosystems worldwide provide abundant evidence of increasing climate impacts to these regions. As mentioned above, temperatures for the Arctic north are rising at two times or more the global mean rate of warming (Screen & Simmonds, 2010). Tree mortality across the Russian boreal forest has increased over the late 20th and early 21st centuries (Allen et al., 2010). The same region has also seen a substantial intensification in fire occurrence, with the fire return interval falling from 101 years in the 19th century to 65 years in the 20th century for larch-dominant forest stands (Kharuk, Ranson, & Dvinskaya, 2008). Increased recurrence of wildfires is reducing the carbon stocks of affected boreal forest sites (Palviainen et al., 2020), altering soil and permafrost regimes (Gibson et al., 2018), changing historic species compositions (Mack et al., 2021), and in some cases leading to post-fire "regeneration failure" (Burrell, Kukavskaya, Baxter, Sun, & Barrett, 2021). Forest area burned has correspondingly increased across Siberia based on data from multiple sources (Soja et al., 2007). The extent of wildfires in boreal environments is widely anticipated to continue increasing in the future (Balshi et al., 2009; Kloster, Mahowald, Randerson, & Lawrence, 2012; Shuman et al., 2017; Wotton, Flannigan, & Marshall, 2017). Satellite data also highlight decreases in forest productivity across Alaska since 1982 (Beck et al., 2011). With regional warming, Alaska has also begun to see large-scale forest mortality events driven by previously cold-limited spruce beetles, as reviewed in (Soja et al., 2007). For example, an ongoing spruce beetle outbreak in south-central Alaska has affected ~0.5 million ha since 2016 (US Forest Service, 2019). The North American boreal forest is also exhibiting an increase in the proportion of deciduous tree cover (J. A. Wang et al., 2019). Replacement of boreal forest by grasslands and deciduous forest is expected to continue based on the response of vegetation models to climate forcing (Foster et al., 2019; J. A. Wang et al., 2019). Together with climate-driven changes to precipitation and soil moisture, increasing insect outbreaks and more frequent and intense wildfires may drive declines in boreal forest productivity within the interior of boreal regions (Boyd et al., 2021; Foster et al., 2019; J. A. Wang et al., 2019). Region-specific modeling studies focusing on Alaska (Foster et al., 2019; Mann, Rupp, Olson, & Duffy, 2012), Siberia (Shuman et al., 2015), and China's boreal forests (Wu et al., 2017) support the potential for future climate-induced shifts in vegetation consistent with such patterns.

Yet while boreal forest productivity and tree cover are on the decline at the southern edge of the boreal zone and within interior regions, thirty-year datasets of satellite and observational evidence also point towards ongoing expansion of boreal forests northwards into area previously occupied by tundra thanks to warmer temperatures (Beck et al., 2011; Ju & Masek, 2016; Pastick et al., 2019; Pearson et al., 2013). Since 1960, the growing season across the boreal zone has lengthened by 3 days/decade (Euskirchen et al., 2006). Expansion of trees into the tundra biome has implications for regional and global climate, as the albedo of forests is lower than that of tundra, leading to warmer winter conditions with greater tree cover (Bonan et al., 1992). The spatial extent and speed of boreal treeline migration is likely to be dependent on a number of factors (Rees et al., 2020), including nutrient and permafrost conditions and wildfire. However, increasing temperatures are likely to drive boreal expansion, leading to cascading impacts on soil, disturbance, and climate regimes.

The accelerated pace of boreal climatic shifts relative to the rest of the world is likely to continue over the 21st century. Warming of 3-5°C globally by end-of-century would imply average temperature increases of 7-10°C for large parts of Russia, with regional warming of up to 12°C (Schaphoff et al., 2016). Such warming would likely considerably exacerbate the abovementioned climate impacts upon the boreal environment. For example, one study predicts that the probability and intensity of Canadian boreal forest fires might more than double across large areas by 2080-2100 under an RCP8.5 scenario (Wotton et al., 2017), while another recent analysis modeled mean potential increases in burned area of 29-35% for the Northwest Territories and 46-55% for interior Alaska by 2050-2074 under RCP8.5, driven predominantly by more frequent occurrence of lightning (Veraverbeke et al., 2017).

The rapid pace of such observed and predicted patterns, which in some cases exceed older predictions, raises the possibility that future change and warming-induced feedbacks within the boreal biome may proceed non-linearly rather than linearly (Foster et al., 2019; Johnstone, Hollingsworth, Chapin, & Mack, 2010; Soja et al., 2007). An extensive survey of forest cover across the boreal environment has indicated that intermediate states of landscape tree cover are rare and potentially unstable, suggesting that forested areas may transition to systems with sparse tree cover more abruptly than previously thought (Scheffer et al., 2012). Shifts towards more prevalent fires potentially play a major role in driving a transition towards more deciduous tree cover (Johnstone et al., 2010). Decades of remote sensing data from North American boreal forests suggest that current CMIP6 models are failing to account for notable impacts from disturbances like fire and insects, while overestimating additional biomass accumulation from CO_2 growth enhancement (J. A. Wang, Baccini, Farina, Randerson, & Friedl, 2021). Paleoclimate evidence of boreal forests responding strongly to temperature changes from natural orbital patterns, coupled with a strong modeled importance of variability in boreal forest area for overall climate, suggests that large shifts in boreal biome extent could act as a large-scale climate feedback and have done so in the geologic past (Foley et al., 1994).

Ecosystem changes for the boreal forest may take place within decades to a century. A multi-region model analysis found that for large portions of all major boreal regions with the exception of Eastern North America, projected temperature changes under worst-case emissions would alter local climatic conditions to resemble current woodland/shrubland climates by 2090 (Gauthier et al., 2015). (Mann et al., 2012) in fact suggest that a transition to mixed woodlands has already begun for the Alaskan boreal region and is anticipated to reach completion by mid-century. Projections for Siberia also indicate a relatively rapid shift, with dark and light-needled forest declining from ~60% of the modeled area to ~40% and ~24% for a 720 ppm and >800 ppm end-of-century emissions scenario, respectively (Shuman et al., 2015).

However, a notable degree of uncertainty continues to surround the critical thresholds, extent of change, and climatic impacts of large-scale boreal ecosystem shifts. Polling of expert researchers produces a wide range of increases in global mean temperature relative to pre-industrial temperature (3-4°C) for intensifying boreal forest transitions (Kriegler et al., 2009). Many modeling studies to date have focused heavily on worst-case emissions scenarios like RCP8.5 (Gauthier et al., 2015; Shuman et al., 2015), which may no longer represent most-likely climate trajectories (Hausfather & Peters, 2020). Comparisons of vegetation shifts under multiple climate scenarios suggest that boreal forest changes may be strongly mitigation-dependent, with more limited impacts for more moderate emissions pathways (Lucht, Schaphoff, Erbrecht, Heyder, & Cramer, 2006). Loss of boreal forest in southern regions may also be partially compensated by afforestation along the northern boundaries of the boreal biome as warmer, less-frozen tundra gives way to forested landscapes (Scheffer et al., 2012). Existing large-scale vegetation models also remain limited in their ability to replicate complex terrestrial ecosystem dynamics and produce confident estimates for changes in biomass carbon (Schaphoff et al., 2016). Finally, the future of boreal forests will also strongly depend on human management practices.

	Boreal forest zone		
	Southern margins	**Boreal zone interior**	**Northern margins**
Warming feedbacks in response to warming	Soil organic matter decomposes at higher rates Increased wildfire impacts Increased boreal tree mortality from pests	Soil organic matter decomposes at higher rates Increased wildfire impacts Increased boreal tree mortality from pests	Soil organic matter decomposes at higher rates Increased wildfire impacts Expansion of boreal conifers into previously bare ground darkens land surface
Cooling feedbacks in response to warming	Carbon storage eventually shifts aboveground into more robust deciduous vegetation Replacement of dark-leaved boreal conifers increases land surface reflectivity	Carbon storage eventually shifts aboveground into more robust deciduous vegetation Replacement of dark-leaved boreal conifers increases land surface reflectivity	Faster vegetation growth due to warmer temperatures stores more carbon
Net effects	Highly uncertain, but with a considerable possibility for net warming due to increased carbon release from decomposition and burning of organic matter in boreal forest soils. Climate impacts may also differ by region (i.e. E. Siberia vs. W. Siberia vs. N. Canada vs. Alaska)		

Table 4: Conceptual outline of changes within different regions of the boreal zone organized by their anticipated impacts upon climate.

2.5.3 The importance of boreal forest regions for global climate

Overall, the potential for boreal forests to exhibit tipping element behavior remains uncertain, but troubling (Table 4). The recent pace of change witnessed in the boreal north clearly points to the potential for large-scale transitions. Projections for such a transition, however, remain based on models that must make simplifying assumptions about complex mechanisms such as precipitation, fire, and soil moisture availability and their effects on needle-leaved trees. These processes as well as differential responses of tree species to environmental changes (Dusenge, Madhavji, & Way, 2020) and competitive interactions between individual trees and species are important factors for modeling the response of boreal regions to climate forcing (Foster et al., 2019; Shugart et al., 2015). Due to landscape and ecological heterogeneity, some boreal areas may prove relatively more resilient to climate change thanks to favorable terrain or ecosystem conditions (Stralberg et al., 2020). Nevertheless, parallel findings from a growing number of analyses predicting major vegetation shifts across the biome region underline the real possibility for a boreal transition beginning later this century.

One of the most worrying aspects of a potential boreal tipping element involves the difficulty of quantifying the climatic impact of replacement of boreal forests by deciduous woodlands or shrublands. Boreal forests contain 30% or more of global soil carbon, and up to 95% of organic carbon within boreal forest ecosystems may be stored belowground (Bradshaw & Warkentin, 2015; Flannigan et al., 2009). Small-scale processes triggered by warming, hydrological changes, and permafrost thaw may precipitate the release of some of this large carbon pool via decomposition (M. R. Turetsky et al., 2019), while fire and post-fire mortality have the potential to emit considerable carbon from both forest biomass as well as upper soil organic layers (Paulo Monteiro Brando et al., 2014; De Groot et al., 2013; Dieleman et al., 2020; Shvidenko et al., 2011). At the same time, projections for potential carbon release from boreal soils are complicated by a high degree of overlap with estimates for permafrost carbon release (Bradshaw & Warkentin, 2015), as the northern boreal and permafrost zones partially coincide. A potential also exists for interactions between permafrost thaw and vegetation shifts. For instance, observations in Canada suggest that permafrost thaw may push boreal landscapes towards wetland-like conditions, resulting in declines in forest cover (Carpino et al., 2018). Warming additionally affects carbon gains and losses in boreal vegetation thanks to shifts in productivity and respiration, with potentially opposing seasonal effects (Z. Liu et al., 2020).

Some research from an Alaskan study suggests that after wildfires, a shift in vegetation from needle-leaved trees to deciduous forest would not substantially change the total pool of organic carbon, and would potentially increase the storage lifetime of carbon due to the greater resistance of deciduous biomass to fire and decay (Alexander & Mack, 2016; Mack et al., 2021). Biotic feedbacks such as vegetation shifts towards hardwoods during post-fire succession may also improve resilience to fire (Hansen, Fitzsimmons, Olnes, & Williams, 2020) and reduce the future frequency of wildfires relative to model projections that predict fire activity from changes in climate alone (Marchal, Cumming, & McIntire, 2020). Apart from implications for fire and carbon storage, a shift in vegetation from darker coniferous species to lighter deciduous trees would increase regional reflectivity of the southern and interior areas of the boreal zone, acting as a negative feedback on warming (R. A. Betts, 2000; Z. Liu, Ballantyne, & Cooper, 2019; Mykleby, Snyder, & Twine, 2017; Piao et al., 2020).

As boreal forests expand northwards into current tundra biomes, however, the same research suggests that the darker surface of boreal vegetation – particularly as it overtakes treeless ground previously snow-covered in winter – will act as a positive feedback reinforcing climate change (R. A. Betts, 2000; Z. Liu et al., 2019; Mykleby et al., 2017; Piao et al., 2020). This albedo-driven effect will likely overcome any positive benefits from added carbon sequestration from northwards expansion of boreal vegetation. Ultimately, current research cannot eliminate the possibility that changes across the boreal zone due to a warming climate could act as a net positive climate feedback, thanks to the role of permafrost thaw and wildfires in liberating soil carbon that makes up the majority of stored carbon across this ecosystem.

Overall, calculations of changes to carbon stocks, regional albedo, carbon sinks, and the timescales involved even at local or regional scales remain imprecise and depend upon multiple complex processes and feedbacks (Foster et al., 2019; Shuman et al., 2015). Consequently, estimating the climatic impact of worldwide changes to the boreal biome under expected future emissions remains challenging. Given the sheer magnitude of carbon sequestered within boreal soils and forest biomass as well as the areal extent of boreal regions, however, boreal forest dieback and shifts represent one of the more potentially immediate and significant climate tipping elements.

2.6 Disruption of tropical seasonal monsoons

2.6.1 Background

Seasonal monsoon precipitation is the primary source of water for nearly all tropical land, governing water availability and agricultural yields for billions of people in much of Africa and Asia, the tropical Americas, and northern Australia (Christensen et al., 2013). Furthermore, extreme precipitation in these regions typically results from storms embedded within the continental-scale monsoon winds, causing flooding and landslides that often inflict significant loss of life and property. Given the importance of monsoons to food security, water availability, and natural disaster risk, researchers worldwide have devoted sizable effort to predicting changes in monsoon seasonality and intensity that might result from anthropogenic climate change.

Although monsoons are sometimes viewed as regional phenomena, they are part of the global circulation of the atmosphere, with warm air rising over tropical continents in the summer hemisphere, flowing across the equator, and sinking as it cools in the winter hemisphere (Fasullo & Webster, 2003). During solstice seasons, the sum of all the regional monsoons constitutes the global Hadley circulation (Webster, 2004). The seasonal cycle of monsoons and the Hadley circulation is driven by the seasonality of solar radiation, with the energy of intense summer sunlight transferred rapidly to the overlying atmosphere due to the low heat capacity of land. The vertical transport of this energy from the thin layer of near-surface air into the bulk of the overlying atmosphere is performed primarily by precipitating clouds, yielding monsoon precipitation (Biasutti et al., 2018). Changes in monsoon precipitation thus involve planetary-scale atmospheric flow, comparatively small-scale clouds, surface and atmospheric absorption of solar radiation, and the properties of land and ocean surfaces.

Because of difficulties in understanding and simulating (e.g. in global climate models) the multi-scale, coupled monsoon systems described above, many studies have turned to the historical record for indications of ongoing shifts in monsoon rainfall; a repeating theme is the identification of large-scale changes in seasonal mean monsoon rainfall in many regions over the last century, with some reversal of those changes in recent decades. For example, monsoon rainfall over central India decreased about 10% between 1950 and 2000, with that trend attributed to various possible causes such as atmospheric aerosols (Bollasina, Ming, & Ramaswamy, 2011; Ramanathan et al., 2005), Indian Ocean warming (Roxy et al., 2015), land use change (Paul et al., 2016), and irrigation (Niyogi, Kishtawal, Tripathi, & Govindaraju, 2010; Shukla, Puma, & Cook, 2014). However, that decrease in central Indian rainfall ended around the year 2000 and the negative rainfall anomaly has since almost completely recovered in most observational records (Q. Jin & Wang, 2017). When averaging over the larger region of the whole country of India, precipitation shows no detectable trend in rain gauge datasets extending back in time more than a century, for both summer and the full calendar year (Saha, Chakraborty, Paul, Samanta, & Singh, 2018). In Africa's Sahel, monsoon-season rainfall decreased about 40% from the 1950 to the mid-1980s, then partially recovered over the next 20 years (I. M. Held, Delworth, Lu, Findell, & Knutson, 2005). Remote SST variations are strongly associated with this decadal variability in Sahel rainfall (Folland, Palmer, & Parker, 1986; Giannini, Saravanan, & Chang, 2003), with possible roles for anthropogenic aerosols and natural ocean-atmosphere variability in causing those SST changes (Biasutti & Giannini, 2006; Rotstayn & Lohmann, 2002). Turning from the seasonal mean to the temporal characteristics of rain events within the monsoon season, studies have found evidence for Sahel precipitation becoming increasingly extreme and erratic (Biasutti, 2019). For example, 35 years of satellite imagery showed an increase in the number of intense mesoscale convective systems just south of the Sahara (Taylor et al., 2017). In central India, some studies have found an intensification of rainfall in wet periods and a reduction in the intensity of dry spells (Singh, Ghosh, Roxy, & McDermid, 2019; Singh, Tsiang, Rajaratnam, & Diffenbaugh, 2014).

This section does not seek to synthesize all historical and projected future trends in monsoon rainfall, but asks whether monsoons constitute a tipping element of the climate system (other reviews, such as (Hoell, Funk, Barlow, & Shukla, 2016; Pascale, Carvalho, Adams, Castro, & Cavalcanti, 2019; A. G. Turner & Annamalai, 2012; Bin Wang, Jin, & Liu, 2020) summarize the regionally disparate observed trends in monsoon rainfall). Is it reasonable to expect abrupt changes in monsoon characteristics in response to anthropogenic climate forcings, and if so would such changes be reversible? Studies have indeed suggested that anthropogenic forcings might cause a sudden reduction in South Asian monsoon strength and an abrupt increase in West African monsoon intensity in the next century (Bathiany, Scheffer, Van Nes, Williamson, & Lenton, 2018; Lenton et al., 2008; Levermann, Schewe, Petoukhov, & Held, 2009; Schewe, Levermann, & Cheng, 2012; K. Zickfeld, B. Knopf V, . Petoukhov, & H. J. Schellnhuber, 2005). However, these studies have relied on highly idealized and simplified sets of differential equations, often distinct from the formal primitive equations of fluid motion that form the basis of global climate models, and their relevance to monsoons has been disputed (Boos & Storelvmo, 2016b; Seshadri, 2017). The majority of research to date, including the ensemble of CMIP5 and CMIP6 models, suggests that the global monsoon domain will experience a gradual increase of seasonal mean precipitation and a gradual weakening of low-level mean winds in response to warming, albeit with many uncertainties and caveats involving model bias (Chen et al., 2020; Christensen et al., 2013; Hill, 2019; O. Hoegh-Guldberg et al., 2018; C. Jin, Wang, & Liu, 2020). Yet, the great importance of monsoons for billions of people living in the tropics justifies continued study of even a slim chance of abrupt change, especially given the existing evidence for some abrupt variations in both paleo monsoons and the seasonal cycle of modern-day observed monsoons. The rest of this section examines the possibility of abrupt changes in monsoons that might occur in response to comparatively steady global or regional forcings, then closes with a very brief summary of projected future trends in regional monsoons that may be unrelated to tipping elements.

2.6.1 Evidence for abrupt transitions in the modern-day seasonal cycle and paleo monsoons

Multiple regional monsoons exhibit threshold-like transitions in early summer, switching abruptly from a dry, winter-like state to a wet summer state. In early summer, precipitation and continental-scale atmospheric flow in the South Asian monsoon intensify more rapidly than can be explained by a linear response to the solar forcing (Boos & Emanuel, 2009; D. Halpern & Woiceshyn, 1999; Krishnamurti, Ardanuy, Ramanathan, & Pasch, 1981; Murakami, Chen, & Xie, 1986). The transition from wintertime trade winds (directed westward) to eastward monsoon flow occurs rapidly--over a time scale of a few days--in the Australian monsoon, although with perhaps greater association with intrinsic intraseasonal variability and less association with the solar forcing than other monsoons (Wheeler & McBride, 2005). Convective cloud activity over the Sahel shifts rapidly from 5ºN to 10ºN during the onset of the West African monsoon (Fontaine, Gaetani, Ullmann, & Roucou, 2011; Sultan & Janicot, 2003).

Given this observed behavior in the seasonal cycle, is it reasonable to expect that monsoons might abruptly shift into a much drier or wetter state as a long-term climate forcing is applied? If one could reduce the effective solar forcing below the level needed to produce the observed seasonal onset of the summer monsoon in a region, say by increasing the concentration of reflective atmospheric aerosols, one would expect the summer monsoon to fail to begin. However, such a forcing would need to be very large: insolation increases by about 250 W m^{-2} between winter and summer solstices at 30ºN, while regional aerosol radiative forcings typically peak around 10 W m^{-2} or less (Ramaswamy et al., 2001; Takemura, Nozawa, Emori, Nakajima, & Nakajima, 2005). Changes in radiative forcing due to historical changes in land use and land cover, including past deforestation and cropland expansion, is of similarly small magnitude (C J Smith et al., 2020). Furthermore, monsoons have more degrees of freedom than an idealized mathematical step function, and are known to exhibit changes in intensity, onset and withdrawal dates, and spatial structure in response to a variety of forcings. For example, the well-known reduction in Indian monsoon rainfall that occurs during El Niño events has been shown to be associated with a shortening of the rainy season (Goswami &

Xavier, 2005), and one of the most common patterns of interannual variability in the West African monsoon consists of a north-south shift of the rainfall maximum (Nicholson & Grist, 2001). Some of the simple models used to argue that an abrupt failure of monsoons will occur in response to climate forcings can only represent changes in monsoon intensity, and fail to include such well-observed spatial shifts or changes in duration (Bathiany et al., 2018; Levermann et al., 2009).

Paleoclimate proxies provide abundant evidence for abrupt changes in annual mean rainfall in several monsoon regions, although the meaning of "abrupt" and the drivers of these changes require careful consideration before they can be used as analogues for future climate change. These proxies include not only stable isotopes, but easier to interpret quantities such as pollen concentrations, lake levels, sediment deposition rates, and macrofossils (COHMAP Members, 1988). The West African monsoon is a well-studied example, with many proxies indicating a step-like beginning and end to the African Humid Period, which comprised a large expansion of rainfall across the Sahara concurrent with the changes in Earth's orbit that brought more intense boreal summer insolation roughly 6 000 to 15 000 years ago (deMenocal et al., 2000; Gasse, 2000). Compared to some anthropogenic forcings, such as increases in greenhouse gas concentrations which exert an order 4 W m^{-2} radiative forcing that is relatively uniform in space, this mid-Holocene insolation anomaly was large in magnitude and non-uniform, peaking at 30 W m^{-2} near the North Pole but absent at the South Pole, thus likely more efficiently driving thermally direct circulations such as monsoons by enhancing the meridional gradient of insolation. This insolation forcing amplified and then decayed over thousands of years, while proxy time series show North African hydrology responding over centuries or even decades, which is comparatively fast (Adkins, deMenocal, & Eshel, 2006). However, it remains unclear whether these potentially abrupt changes manifested simultaneously over most of North Africa or occurred sequentially at different latitudes as the narrow summer-mean monsoon rain band gradually shifted north-south in response to the insolation forcing. (McGee, deMenocal, Winckler, Stuut, & Bradtmiller, 2013) found that past changes in dust deposition off Africa's west coast were consistent with synchronous changes in dust across all of northern Africa, suggesting a truly rapid expansion and contraction of the West African monsoon. In contrast, (Shanahan et al., 2015) found that hydrogen isotopes in leaf waxes indicated precipitation changes that were only locally abrupt at the end of the African Humid Period, occurring at progressively later times at lower latitudes as the rain belt presumably withdrew gradually toward the equator.

Potentially abrupt changes exist in the proxy records of other paleo monsoons, with related questions regarding locality and relevance for future climate change. For example, high-resolution time series of isotopes in cave deposits extend the record of the East Asian monsoon hundreds of thousands of years into the past, showing step-like responses over many cycles of the smooth, sinusoidal insolation forcing (Y. Wang et al., 2008). However, the East Asian monsoon is distinct from the tropical monsoons discussed in this section, being governed by midlatitude jet stream dynamics (Liang & Wang, 1998; Sampe & Xie, 2010); changes in cave isotopes in East Asia may furthermore indicate changing isotopic fractionation in remote regions thousands of kilometers away, or changes in the source regions of water vapor fluxes (J.-E. Lee et al., 2012; Maher, 2008; Pausata, Battisti, Nisancioglu, & Bitz, 2011). Another example is the abrupt change in the South Asian monsoon argued to have occurred in response to geologic uplift of the Tibetan Plateau, with the forcing evolving over millions of years and much debate about the physical meaning of proxies (Molnar, Boos, & Battisti, 2010).

2.6.2 Proposed mechanisms for abrupt changes in monsoons

Although mechanisms for the abrupt seasonal onset of monsoons have been studied for decades (Numaguti, 1995; Plumb & Hou, 1992; Xie & Saiki, 1999), most recent arguments for monsoons being a tipping element of the climate system have focused on a particular moisture-advection feedback (K. Zickfeld et al., 2005). In this feedback, water vapor is transported from ocean to warmer and drier land in the thermally direct monsoon circulation; as the water vapor condenses in rising air over land, the latent heat release reinforces the existing land-sea temperature contrast, strengthening the circulation in a self-reinforcing dynamic (Levermann et al., 2009). Anthropogenic aerosol emissions or land use shifts could increase regional albedo, causing cooling over land that weakens the temperature gradient and disrupts the moisture-advection feedback, causing the monsoon to abruptly shift into a dry state (K. Zickfeld et al., 2005). This tipping point was proposed to exist at a regional albedo of 0.5, with land use change and aerosol emissions over South Asia possibly increasing albedo to that threshold (Lenton et al., 2008; K. Zickfeld et al., 2005). Studies have explored potential humidity thresholds for this feedback, arguing that past variations in humidity over ocean allowed the feedback to cause abrupt changes in paleo monsoons (Schewe et al., 2012).

The existence of this moisture-advection feedback on monsoon strength has been disputed, because the simple box model used to mathematically formulate the feedback omitted the fact that atmospheric temperature drops as air rises along an adiabat (Seshadri, 2017). Other authors made a consistent criticism (Boos & Storelvmo, 2016a), noting that the moisture-advection feedback model omits the static stability of the troposphere; adding that stabilizing term to the model equations resulted in a continuous and nearly linear response to forcings. Although (Levermann, Petoukhov, Schewe, & Schellnhuber, 2016) asserted in response that this static stability is insufficient to counteract the moisture-advection feedback, pointing to paleoclimate evidence of abrupt monsoonal transitions, static stability has been shown in decades of research to balance high-amplitude atmospheric heatings (Boos & Storelvmo, 2016b), and the occurrence of abrupt shifts in paleo monsoons does not imply the existence of a moisture-advection feedback (see previous subsection). Setting aside arguments about the formulation of simple box models, (Boos & Storelvmo, 2016a) showed that, in an ensemble of global climate model simulations, no abrupt transitions, hysteresis, or other behavior characteristic of tipping points occurred when the South Asian monsoon was subject to a wide range of greenhouse gas, surface albedo, and aerosol forcings. Furthermore, the proposed moisture-advection feedback for abrupt monsoon transitions is similar in formulation to the moisture-convergence feedback proposed as the cause of hurricanes in the 1960s (Jule G Charney & Eliassen, 1964) and later shown to fail in producing realistic simulations of those storms by unrealistically producing the most rapid amplification at the smallest length scales (Yanai, 1964). A review of tropical atmospheric dynamics called the idea that low-level moisture convergence causes precipitation and latent heating “an influential and lengthy dead-end road in atmospheric science” (Emanuel, David Neelin, & Bretherton, 1994). This does not mean that atmospheric moist dynamics are incapable of causing nonlinear, abrupt changes: (Dixit, Sherwood, Geoffroy, &

Mantsis, 2018) used idealized heatings in a global climate model to argue that dry air advection from the Sahara suppresses West African rainfall in the modern climate, and that a northward shift of that rainfall in response to the mid-Holocene insolation forcing might have shut down that dry air advection and produced nonlinear intensification of the West African monsoon. (Seshadri, 2017) found bifurcations indicative of tipping points when varying the parameters of a simple box model of monsoons, but noted that it was important to determine whether those parameter values were realistic.

Although recent discussion of monsoon tipping points has focused on a moisture-advection feedback (Lenton et al., 2008), the most well-known mechanism for abrupt monsoon change may be the desert-albedo feedback proposed by (J G Charney, 1975) in the context of intense drought in the Sahel that started in the late 1960s. That feedback is biogeophysical in nature, with reduced rainfall causing drying and vegetation destruction over land with a consequent increase in land surface albedo, which in turn reduces absorbed sunlight and thus the radiative forcing for monsoon rainfall (J G Charney, 1975). This idea led to the Sahel and other drylands being thought of as fragile systems prone to nonlinear feedbacks, with modest degradation in vegetation cover prone to push those systems into a new, desert equilibrium (Biasutti, 2019). However, observed association of the Sahel drought of the 1970s and 1980s with decadal variations in SSTs, together with global climate model simulations, led to remote ocean temperature variations now being widely regarded as the cause of that drought (Folland et al., 1986; Giannini et al., 2003). Feedbacks between vegetation and precipitation are now regarded as a possible amplifier of the remote oceanic forcing of the 1970s-1980s drought and are thought to be critical for producing the vast expansion of precipitation across the Sahara during the African Humid Period roughly 6 000 years ago (Biasutti, 2016, 2019; Boos & Korty, 2016). Quantitative understanding of such biogeophysical feedbacks remains a research frontier (National Research Council, 2013), with parameterization of the vegetation component in climate models being a major area of research and development (Fisher et al., 2018). Although some have argued that a feedback between vegetation and precipitation may create a tipping point in the West African monsoon that would cause grasslands to reduce the area of the Sahara desert at a rate of 10% per decade (Claussen, Brovkin, Ganopolski, Kubatzki, & Petoukhov, 2003; Lenton et al., 2008), the state of both knowledge and numerical model representations of such feedbacks provides low confidence in such claims.

2.6.3 Monsoon response to abrupt changes in other earth systems

Monsoons might undergo abrupt changes not because of their own internal, nonlinear dynamics, but because they are forced by large amplitude, abrupt changes in some other element of the Earth system. A canonical example is the abrupt variation of the African and Asian monsoons during the last ice age, thought to occur in response to the rapid temperature changes in the North Atlantic that are known as Heinrich and Dansgaard-Oschger events (REF). Large discharges of fresh ice from the Laurentide ice sheet into the Atlantic are hypothesized to have slowed the ocean circulation and cooled the entire northern hemisphere, drying and weakening the northern summer monsoons (REF). In this scenario, monsoons may be responding predictably and even linearly to an abrupt forcing. Such scenarios bear important lessons for the possible response of monsoons to abrupt changes in the Greenland or Antarctic ice sheets, the Atlantic Meridional Overturning Circulation, or coastal stratocumulus cloud decks.

2.6.4 Synthesis and more general projections of monsoon change

Overall, mechanisms for abrupt monsoon regime change in response to global warming remain a topic of active discussion, in part because such changes, even if extremely unlikely, could have catastrophic societal impact. Nevertheless, definitive evidence for the abrupt response of monsoons to gradual forcings remains elusive. This is not to imply that large, gradual, and high-impact changes in monsoons are unlikely to occur; many analyses of the CMIP model projections that are performed periodically for IPCC reports indicate coming changes in many regional monsoons. In the latest collection of such models (CMIP6), there is general agreement that annual mean rainfall will very likely increase over most of the Asian and African monsoon domains over the next century, but will likely decrease in the core North American monsoon region (C. Jin et al., 2020; Bin Wang et al., 2020). Future changes in South American and Australian monsoon precipitation are close to zero in these models. Projected changes in seasonal-mean precipitation generally increase in magnitude with the strength of the greenhouse gas forcing (Chen et al., 2020; C. Jin et al., 2020), providing no obvious indication of strongly nonlinear or tipping element behavior in the models. Subseasonal variability in precipitation is also expected to increase in nearly all monsoon regions, with an increase in the intensity and frequency of precipitation extremes accompanying enhanced drought risk in some areas (Bin Wang et al., 2020). Such changes will have large socio-economic impact even in the absence of tipping points.

2.7 Catastrophic warming due to breakup of stratocumulus cloud decks

2.7.1 Background

Stratocumulus cloud decks are a distinct feature over subtropical ocean. They cover some 20% of tropical oceans (Eastman, Warren, & Hahn, 2011) and play an important role in the global energy balance. These wide swaths of cloud cover reflect 30-60% of incoming shortwave solar radiation (Wood, 2012), in contrast to the high absorptivity of exposed open ocean waters. Accurate representation of cloud dynamics have remained a weakness of large-scale climate models, resulting in uncertainties arising from parameterizations of cloud processes (J. L. Lin, Qian, & Shinoda, 2014; Nam, Bony, Dufresne, & Chepfer, 2012). Such challenges have persisted despite the considerable importance of clouds to Earth's radiative balance.

A recent scientific paper has proposed that under very high CO_2 emissions scenarios (1200 ppm CO_2 equivalent), an extreme climate feedback may occur in which stratocumulus cloud decks disintegrate, triggering rapid and substantial warming of 8°C globally (Schneider, Kaul, & Pressel, 2019). Stratocumulus cloud decks are maintained by a key set of temperature and moisture-driven mechanisms that are in turn affected by atmospheric radiative forcing and surface warming. Stratocumulus cloud deck breakup has been proposed to occur when greenhouse gas-induced weakening of radiative cooling at the top of cloud layers in combination with elevated fluxes of warm, dry air across the temperature inversion in the upper troposphere disconnects cloud decks from their surface moisture supply.

This extreme climate feedback remains an emerging theory, presented in just a couple research publications to date. However, the underlying principles of the mechanisms involved, in which weakened radiative cooling decouples clouds from moisture inputs, are well-established (Bretherton & Wyant, 1997), providing theoretical grounding for this hypothesis. Cloud feedbacks remain a poorly-constrained yet highly influential component of climate models, with researchers stressing clouds' relatively fast response to CO_2 forcing and rising temperatures could potentially drive large changes in radiative forcing (Caballero & Huber, 2013). The stratocumulus cloud deck evaporation mechanism is suggested to only occur at concentrations of CO_2 that are not anticipated to occur this century even under worst-case emissions pathways. At any rate, the significant and rapid temperature spike that would accompany stratocumulus cloud deck evaporation represents a catastrophic outcome that strongly justifies greatly increased research attention to this potentially abrupt tipping element.

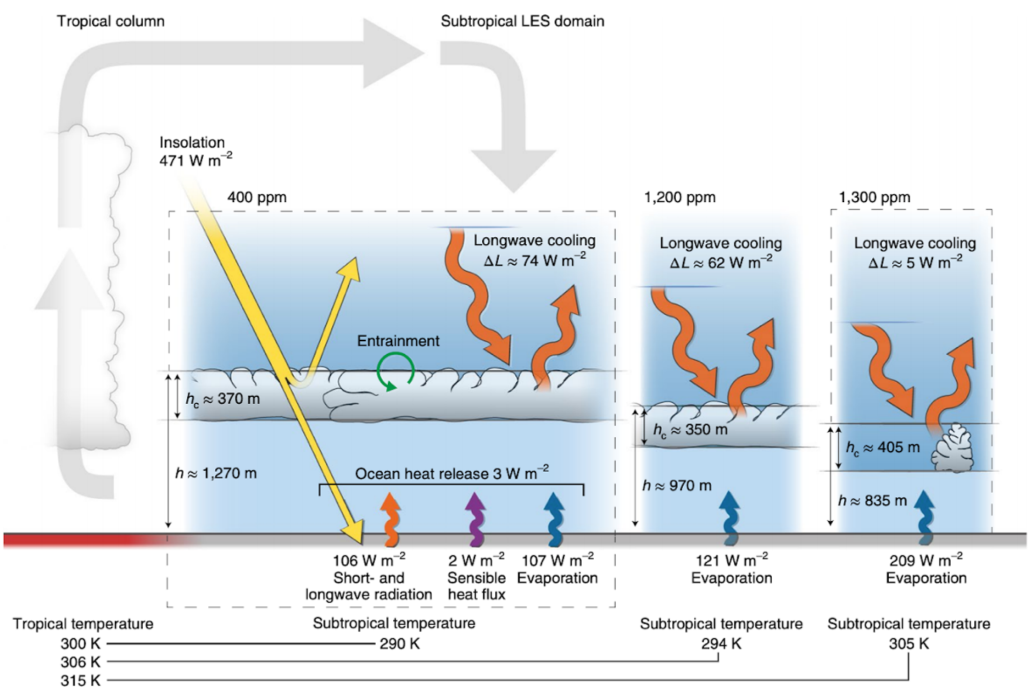

Figure 5: Schematic diagram illustrating mechanisms of stratocumulus cloud deck collapse and resulting consequences upon net radiative energy balance. Under present-day conditions (400 ppm), stratocumulus cloud decks are sustained by longwave radiative cooling at their topmost extents, which drives convective circulation that resupplies moisture to the cloud layer from the ocean's surface. Increasing greenhouse gas concentrations (1200 ppm) increase the input of longwave radiation to stratocumulus clouds from above, reducing the strength of longwave cooling. Beyond a certain threshold of greenhouse gas levels (1300 ppm), longwave

cooling weakens to the point where cloud decks are cut off from surface moisture, leading to their disintegration and abrupt, acute surface warming from increased absorption of sunlight. Figure originally published in (Schneider et al., 2019)

2.7.2 Mechanisms of stratocumulus cloud deck evaporation

Stratocumulus cloud decks over the subtropical ocean are maintained thanks to a self-sustaining feedback in which emission of longwave radiation from cloud tops plays an important role (Figure 5). Clouds have a high ability to absorb longwave radiation relative to free air. Consequently, re-emission of longwave radiation from cloud tops cools stratocumulus cloud banks from above, fueling convective circulation that supplies these cloud formations with moisture from near-surface air over the ocean (Wood, 2012).

Using a high-resolution large-eddy simulation of a representative summertime subtropical ocean region, (Schneider et al., 2019) determined that a sufficiently high concentration of atmospheric greenhouse gases could induce rapid disintegration of stratocumulus cloud decks by upsetting this self-sustaining feedback, further suggesting that this process may occur over short timescales at a global scale.

A high greenhouse gas concentration increases the opacity of the air above stratocumulus cloud decks to longwave radiation. This effect increases the downwelling longwave radiative flux from the overlying atmosphere, reducing the difference between this flux and the stronger upwelling radiative flux from the cloud heights. Radiative cooling at the cloud tops weakens with increasing CO_2 concentrations, eventually losing the strength required to drive sufficiently strong convection of air parcels all the way to the ocean's surface.

At the same time, increased surface evaporation due to GHG-induced warming strengthens atmospheric turbulence generated by latent heat release from condensation in clouds. Increased turbulence in turn drives greater entrainment of warm, dry air across the tropospheric temperature inversion (the altitude at which temperature stops decreasing with increasing height in the lower atmosphere, beyond which temperatures warm with altitude). This second effect also serves as a feedback that works against the existence of stratocumulus decks at higher CO_2 concentrations.

Together, these two factors are responsible for breakup of stratocumulus cloud decks in nature (Bretherton & Wyant, 1997), and both mechanisms are expected to intensify with continued CO_2 emissions. (Schneider et al., 2019) found that total cloud deck collapse occurred at modeled CO_2 equivalent concentrations of more than 1,200 ppm, with continuous cloud banks becoming replaced by scattered cumulus clouds. Breakup of stratocumulus cloud decks removes the majority of cloud cover over these ocean regions, allowing greatly increased absorption of solar radiation at the water's surface and triggering substantial, rapid surface warming. The authors calculated that the resulting change to the Earth's overall energy balance would be sufficient to increase surface temperatures worldwide by 8°C, with particularly intense warming of up to 10°C in subtropical regions.

Model runs suggested that cloud deck collapse occurs suddenly rather than gradually. At lower CO_2-eq concentrations of 400-800 ppm, cloud cover remained dense at present levels, albeit with reduced water content. Once collapse was initiated, however, the process proceeds aggressively and rapidly as the abovementioned feedbacks rapidly self-amplify with the loss of cloud cover, likely starting from the edges of stratocumulus regions where stratocumulus cloud decks exist closer to the critical stability threshold. However, because cloud decks in different regions do not share the same proximity to the stability threshold, it remains unclear how abrupt a process cloud deck breakup would be at a global scale.

The authors do note that the precise CO_2 concentration at which stratocumulus cloud deck evaporation is triggered also depends on large-scale atmospheric dynamics that may themselves shift with climate change. They further highlight a potential mediating mechanism reducing the likelihood of collapse in which predicted weakening of large-scale subsidence in the troposphere with continued warming may somewhat buttress the effect of increasing cloud deck instability (Blossey et al., 2013; Bretherton, 2015; Isaac M. Held & Soden, 2006; Tan, Schneider, Teixeira, & Pressel, 2017).

2.7.3 Potentially extreme impacts of this cloud feedback underline need to fill knowledge gaps

The consequences of stratocumulus cloud deck disintegration triggering a highly abrupt warming of 8°C or more would have devastating implications for human society and natural ecosystems worldwide. Such a magnitude of warming would severely challenge humanity's adaptive ability by massively disrupting global agriculture, altering weather patterns, and significantly accelerating sea-level rise, among other effects—all within an extremely short timeframe. A global mean temperature increase of this level would also carry a high risk of activating and exacerbating many of the other tipping elements covered in this review, resulting in further amplified climate impacts.

The Schneider et al. paper further modeled a lagging, hysteresis-like recovery pattern for stratocumulus decks in which these cloud banks only re-form after CO_2 levels drop significantly below the original triggering threshold. Reformation of stratocumulus cloud decks only occurred after CO_2 concentrations fell to nearly pre-industrial levels (<300 ppm CO_2 -eq). This suggests that once stratocumulus cloud deck collapse occurs, the process of restoring them through carbon sequestration, solar geoengineering, or other means may involve an extremely delayed recovery. At the same time, a subsequent analysis by the same research group suggests that even sustained large-scale solar geoengineering would not fully prevent stratocumulus cloud deck evaporation from taking place, merely raising the requisite critical threshold to 1700 ppm CO_2-eq (Schneider, Kaul, & Pressel, 2020).

A substantial mitigating consideration, however, is that the greenhouse gas concentrations required to initiate stratocumulus cloud deck evaporation (~1200 ppm CO_2 equivalent) are high, at the upper limit of potential greenhouse gas concentrations in 2100 under the very-high RCP8.5 emissions scenario. With the RCP8.5 pathway itself representing a “worst-case” emission scenario that is unlikely to be realized (Hausfather & Peters, 2020), it appears uncertain that the concentration thresholds required to trigger the stratocumulus cloud deck phenomenon will be reached this century. However, potential uncertainty regarding critical CO_2 thresholds reinforces the need for additional research into this mechanism and further highlights the importance of aggressive climate mitigation efforts.

Overall, stratocumulus cloud deck evaporation should be assessed as a high-impact, potentially abrupt tipping element that emerges as a concern only for greenhouse gas forcing at the upper limits of worst-case emissions scenarios for this century. This hypothesized phenomenon remains a novel problem that has only recently been brought to the attention of the scientific community. The response of stratocumulus cloud decks also represents just one of several potential interactions between clouds and climate forcing, including Hadley circulation-driven changes to cloud cover and shifts in midlatitude storm tracks (Bender, Ramanathan, & Tselioudis, 2012; Caballero & Huber, 2013). Substantial further research efforts are required in order to investigate this potential climate feedback in detail, assess risks and potential impacts, and better clarify the conditions that may trigger cloud deck disintegration or other large-scale changes in cloud forcing.

3 Regional candidate tipping elements

3.1 Continued die-off of tropical, shallow-dwelling coral reefs

3.1.1 Background

Coral reefs are among the most productive and ecologically diverse marine ecosystems worldwide, with their biodiversity collapse potentially driving further uncertainty and unanticipated shifts in regional marine ecology, function and resources. Strong scientific evidence points towards critical temperature tipping points beyond which tropical coral reefs undergo severe ecosystem shocks. Shallow, tropical corals demonstrate high sensitivity to temperatures outside of their accustomed range, and are additionally under stress due to increasing trends in overexploitation of resources, land-based pollution (i.e., effluence, agricultural run-off, and microplastics), ocean acidification, and deoxygenation (Bindoff et al., 2019). Temperature anomalies are a very strong predictor of coral bleaching (Sully, Burkepile, Donovan, Hodgson, & van Woesik, 2019), a phenomenon in which the obligate symbiosis between corals and its photosynthetic dinoflagellate microalgae decouples. As these photosynthetic dinoflagellates of the *Symbiodiniaceae* family provide corals with up to 90% of their energy requirements, the loss of these symbionts for a prolonged period can lead to starvation, reproductive impairment, susceptibility to diseases, and death.

While tropical corals undergo daily environmental fluctuations and have survived substantial evolutionary pressures over the course of geologic history, the current rate of ecosystem change presents a serious threat. As marine heatwaves affect tropical coral reefs worldwide (Baker, Glynn, & Riegl, 2008; Heron, Maynard, Van Hooidonk, & Eakin, 2016; M. D. Spalding & Brown, 2015), coupled with the increasing rate of mean ocean temperature rise (Bindoff et al., 2019), the outlook for corals appears grim. Since the 1980s, researchers have documented three large global-scale coral bleaching events (1998, 2002, and 2015-2016) along with more frequent localized events in response to particularly high sea temperature anomalies, with a fourth global bleaching event likely to occur within the next decade or two (Hughes et al., 2017).

Even for a relatively moderate warming scenario, reef organisms would need to demonstrate the ability to acclimatize and adapt to a temperature change of more than 2°C by the late 21st century, a pace of adaptation likely too extreme for these ecosystems to match (Donner, Skirving, Little, Oppenheimer, & Hoegh-Guldberg, 2005; Laufkötter, Zscheischler, & Frölicher, 2020; Putnam, 2021). Coral researchers consequently show strong agreement that the majority of tropical coral habitats will see sharp declines in biodiversity within the coming decades (Bindoff et al., 2019; Descombes et al., 2015).

The economic and societal impacts of coral reef loss to Indo-Pacific, Caribbean, and other communities can be expected to be substantial as an estimated 500 million people rely on coral reef ecosystems. Coral reefs serve as important factors for fishery productivity, hold cultural significance, and provide shoreline fortification from coastal erosion (Ferrario et al., 2014; Storlazzi et al., 2019). Many small island nations have depended upon economic activity provided by coastal tourism, which existing reef ecosystems play a strong role in attracting (M. Spalding et al., 2017; Weatherdon, Magnan, Rogers, Sumaila, & Cheung, 2016). Degradation of warm-water coral ecosystems thus represents an acute threat to island nations and coastal communities across the Asia-Pacific region, endangering their economic security, access to food, and resilience to sea-level rise and extreme weather.

3.1.2 Ongoing and projected temperature stresses and amplified ocean acidification will drive major loss of coral reefs within this century

Coral bleaching events in response to elevated temperatures have been repeatedly documented (Hughes et al., 2017; Wernberg et al., 2015). Four to six week-long periods where water temperature remains +1°C warmer than the average long-term summer maximum can cause bleaching and high mortality (Jokiel & Coles, 1990). Significant long-term declines in coral abundance have already occurred in tropical coral regions, including the Caribbean (Gardner, Côté, Gill, Grant, & Watkinson, 2003; J. B. C. Jackson, Donovan, Cramer, & Lam, 2014), the Indo-Pacific region (Bruno & Selig, 2007), and the Great Barrier Reef (De'Ath, Fabricius, Sweatman, & Puotinen, 2012) due to overfishing, hurricanes, and severe bleaching events. The severity of bleaching events is further observed to be increasing over time, becoming the most prevalent threat to coral reefs (De'Ath et al., 2012; Hughes et al., 2018).

Coral reef ecosystems have demonstrated a limited ability to adapt to temperature stress and rebound once conditions improve (Brown, Dunne, Goodson, & Douglas, 2002; Coles & Brown, 2003; Jury & Toonen, 2019), with observations of subsequent coral bleaching events being triggered at higher temperatures than previous bleaching episodes (Ritson-Williams & Gates, 2020; Sully et al., 2019). Coral reef ecosystems may recover following mass mortality events, replacing the loss of coral cover and diversity in a decade or more (M. D. Spalding & Brown, 2015). However, further warming is strongly anticipated to outpace the adaptive capacity of corals and time required for recovery (Frieler et al., 2013; O. Hoegh-Guldberg et al., 2007; Hughes, Kerry, Baird, et al., 2019; Veron et al., 2009). Furthermore, the capacity to adapt may be limited to just a small subset of coral species and associated marine organisms, severely thinning the biodiversity commonly found in coral reef ecosystems (Heinze et al., 2015; Hughes, Kerry, Connolly, et al., 2019; van der Zande et al., 2020).

The pace of current and future ocean warming will intensify pressures on corals. Over the past 30 years, global mean sea surface temperatures have risen by 0.015°C per year, with this rate projected to accelerate to an average of 0.027°C per year between 1990 and 2090 (Bopp et al., 2013). Under predicted temperature increases, modelers anticipate bleaching occurring as often as annually or biannually for most reef ecosystems within 30-50 years (Donner et al., 2005; Laufkötter et al., 2020). The potential for warmer waters to alter hurricane intensity and frequency also carries implications for coral ecosystems that can suffer physical destruction, particularly if already under physiological stress (Steneck et al., 2019).

Ongoing ocean acidification represents another important stressor, with mean surface ocean pH declining at a rate of 0.018 units per decade across 70% of ocean biomes (Lauvset, Gruber, Landschützer, Olsen, & Tjiputra, 2015). The IPCC assigns near-certainty to the likelihood of continued surface ocean acidification, with a decline in pH of 0.287-0.29 pH units by 2081-2100 relative to 2006-2015 under the RCP8.5 emissions pathway (Bindoff et al., 2019). Even under the more optimistic RCP2.6 scenario, marine pH declines over the same period by 0.036-0.042. Lower pH requires higher energetic costs for coral to build their calcium carbonate skeleton, while also subjecting dead corals that serve as the foundation for living reef structure to accelerated dissolution and erosion (DeCarlo et al., 2015; Silbiger, Guadayol, Thomas, & Donahue, 2014). Numerous observational studies have confirmed the ongoing negative impacts of ocean acidification upon reef habitats and project continued degradation of corals with further increases in atmospheric CO_2 concentration (Bove et al., 2019; Jiang et al., 2018; Kroeker, Kordas, Crim, & Singh, 2010; Mollica et al., 2018; Orr et al., 2005).

Simultaneously, sea-level rise presents an additional threat to shallow-dwelling reefs through inundation, as waters rise faster than the reefs can grow upwards towards light (Perry et al., 2018). Many coral reefs are also directly subject to human-induced negative impacts such as increased sediment runoff, chemical pollution, nutrient-fueled harmful algal blooms, and damage from fishing activities including dynamite fishing (L. Burke, Reytar, Spalding, & Perry, 2011; B. S. Halpern et al., 2015; Hodgson, 1999). Nutrient pollution (DeCarlo et al., 2020; Donovan et al., 2020), competition with algae (Anton et al., 2020), and marine de-oxygenation also pose threats to reef systems (Bindoff et al., 2019).

Research results demonstrate strong agreement that severe coral reef degradation will likely continue throughout the current century even under optimistic climate mitigation scenarios, with coral abundance declining to 10-30% of today's levels even with warming limited to 1.5°C (O. Hoegh-Guldberg et al., 2019). Warming of 2-2.5°C, well within projections for the 21st century under current emissions rates, would cross tipping points, eliminating >99% of tropical corals and irreversibly transforming coral reef ecosystems (Frieler et al., 2013; O. Hoegh-Guldberg, 2014; O. Hoegh-Guldberg et al., 2019; Ove Hoegh-Guldberg, 2014; Hughes et al., 2017; Carl Friedrich Schleussner et al., 2016).

Currently, a future in which tropical corals undergo significant ecosystem transitions represents the most likely outcome in the absence of extremely aggressive emissions mitigation efforts that limit warming to 1.5°C or less. Although tropical corals may persist in some form, with limited localities or regions potentially serving as refugia from climate forcing (Guest et al., 2018; Mies et al., 2020; Wyatt et al., 2020), chances are high that future coral ecosystems may appear completely unrecognizable compared to their state today. Given the considerable sensitivity of reef habitats to even modest temperature increases under strong climate mitigation scenarios, combined with the impact of multiple stressors like land-based pollution and ocean acidification, avoiding a future in which tropical coral reefs degrade may no longer be possible.

3.2 Loss of Amazon rainforest and conversion of significant rainforest area to savanna-like vegetation

3.2.1 Background

The Amazon region of South America contains the world's largest expanses of old-growth tropical rainforest. This large ecosystem possesses global importance, alone producing an estimated 15% of total terrestrial photosynthesis worldwide (Field, Behrenfeld, Randerson, & Falkowski, 1998). The rainforest region possesses rich biodiversity and also represents a major terrestrial biological carbon sink of some 0.4 – 0.6 Gt C per year (Malhi et al., 2006; Y. Pan et al., 2011), a flux of similar magnitude to the ~0.5-0.6 Gt C per year sequestered by all northern boreal forests globally (Y. Pan et al., 2011; Schaphoff et al., 2013). The Amazon rainforest additionally contains a substantial amount of stored carbon in biomass and soil organic matter, tentatively estimated to range between 150-200 Gt C (Cerri et al., 2007; Gibbs, Brown, Niles, & Foley, 2007; Malhi et al., 2006; S. S. Saatchi et al., 2011). Between all of these factors, the Amazon rainforest represents a critical component of the global carbon cycle.

As with many of the world's major rainforests, a significant fraction of precipitation over the Amazon rainforest is recycled water originating from the rainforest vegetation itself. While on average recycled water accounts for some 25-35% of total precipitation in the Amazon region (Da Rocha et al., 2009; Eltahir & Bras, 1994; Salati, Dall'Olio, Matsui, & Gat, 1979; Zeng, Dickinson, & Zeng, 1996), the contribution of recycled water to total rainfall increases during the dry season. In the dry season, recycling is the majority source of regional water vapor, and also plays a role in initiating the onset of the wet season (W. Li & Fu, 2004; Wright et al., 2017). Consequently, dry season precipitation and length for the Amazon region are directly linked to the total area and health of the rainforest system itself. It has been demonstrated that the Amazonian forest's photosynthesis depends on rainfall (13% to 29% of the forest, located in the south and western parts of the basin), where rainfall is lower than 2000mm yr^{-1}, solar radiation (70% to 76% of the forests, located in central and eastern parts of the basin) and by both in some forest areas (Bertani, Wagner, Anderson, & Aragão, 2017; F H Wagner et al., 2016; Fabien Hubert Wagner et al., 2017). Despite predictions that an increase in atmospheric CO_2 concentrations might increase forest carbon uptake (Lapola, Oyama, & Nobre, 2009), a modelling effort to assess climate change impacts in an eastern forest with no water limitation demonstrated that increasing temperature is a key element linked with a reduction in forest growth (Aubry-Kientz, Rossi, Cornu, Wagner, & Hérault, 2019). Moreover, extreme droughts, which are predicted to be more intense in the tropics (IPCC, 2013), have driven increased tree mortality (Phillips et al., 2009, 2010), and exert long-lasting and cumulative effects on forest photosynthetic capacity (Anderson et al., 2018; S. Saatchi et al., 2013).

In addition to pressure from climatic stress factors, these forests face a more imminent threat. Currently, the Amazon rainforest is experiencing a high rate of deforestation, having lost close to 20% of its pre-1970 area extent to date. One estimate calculates a net carbon loss from aboveground biomass in the Amazon of 5 Gt C between 1996 and 2017 as a result of deforestation and disturbance (Bullock & Woodcock, 2020). At its peak around 2005, deforestation rates approached 30 000 sq km of cut or burned forest per year, and while this fell to an average of 6 000 sq km lost per year between 2011-2015 and has remained low relative to early 2000s rates, current rates of loss nevertheless remain highly unsustainable (INPE, 2021; Carlos A. Nobre et al., 2016). Recent policy shifts in Brazil have once again intensified deforestation, with over 20 000 km^2 clear cut during 2019-2021. March and April 2021 showed the highest rate of deforestation since 2015/16 for those same months (INPE, 2021). Satellite observations suggest that large areas of the Amazon are also suffering degradation due to smaller-scale disturbances that could potentially drive high carbon loss such as fire, drought, selective logging, and firewood collection (Bullock, Woodcock, Souza Jr., & Olofsson, 2020). For example, it has been estimated that edge effects created by forest fragmentation can contribute with an average of 63 ± 8 Tg C $year^{-1}$, corresponding to one-third of losses from deforestation for the 2001 to 2015 period (Silva Junior et al., 2020) and forest fire gross emissions alone (989 ± 504 Tg CO_2 $year^{-1}$) are more than half as great as those from old-growth forest deforestation during drought years (Anderson et al., 2018). The magnitude of these carbon losses highlight the strong requirement for new policies and commitments to tackle deforestation and climate change.

Some Amazon researchers have warned that vegetation loss could ultimately reduce recycled precipitation inputs across large portions of the Amazon basin to the point where large rainforest trees become subject to increased mortality due to the dry season lengthening (Carlos Afonso Nobre & Borma, 2009). However, changes in the seasonality of rainfall have already been observed. It has been demonstrated that the dry-season length has increased over southern (R. Fu et al., 2013) and western (Espinoza, Ronchail, Marengo, & Segura, 2019) Amazonia, with an observed delay of the dry-season's end. Moreover, the altered atmospheric moisture content of air passing over deforested areas results in less rain production relative to air passing over dense forests, leading to further projected rainfall reductions of 12% and 21% in wet-season and dry season precipitation, respectively, across the Amazon by 2050 (Spracklen, Arnold, & Taylor, 2012). For every percentage point increase in deforestation, the onset of the rainy season becomes delayed by 0.12–0.17 days (Leite-Filho, de Sousa Pontes, & Costa, 2019; Leite-Filho, Soares-Filho, Davis, Abrahão, & Börner, 2021) and 10% of additional forest loss induces a net reduction in annual rainfall of -49.2 ± 11.3 mm (Leite-Filho et al., 2021). Therefore, the rainfall cycle is already changing. As rising numbers of trees die due to water stress (Phillips et al., 2009), further reducing the strength of the regional water cycle, dieback of rainforest vegetation becomes a self-sustaining positive feedback, ultimately causing significant sections of the Amazon rainforest to collapse and potentially shifts towards a savanna-like type of ecosystem (Lyra, Chou, & Sampaio, 2016; Carlos Afonso Nobre & Borma, 2009).

Other authors have outlined alternative pathways and considerations for region-wide ecosystem shifts, suggesting that increased forest degradation by wildfires, selective logging and fragmentation represents a more immediate threat to the rainforest (Aragão et al., 2018; P.M Brando et al., 2019; Silva et al., 2020), leading to a loss of many ecosystem services, including biodiversity and carbon storage. On a medium time-scale, these rainforests could shift towards seasonal forest as opposed to savanna grasslands (Malhi et al., 2009), emphasizing uncertainties in the ability of models to predict seasonal precipitation (Good, Jones, Lowe, Betts, & Gedney, 2013), and highlighting the potential for CO_2 fertilization to compensate for biomass losses (Cox et al., 2013; Huntingford et al., 2013). Such factors could mean that changes for the Amazon forest will be driven by local climate-forced conditions, rather than a biome-wide climate threshold. Nonetheless, field-based experiments have demonstrated that abrupt and unpredictable changes may occur: it has been observed an increase of more than 200% in fire-induced tree mortality during a severe drought event, when high temperatures and fuel loads were higher than the long-term average, leading to a decline in canopy cover from 23% to 31%, AGB from 12% to 30% and increase in grasses in the forest edges (Paulo Monteiro Brando et al., 2014). During 2007, a year with positive temperature anomalies, more than 11 000km^2 of forests burned in the Amazon, during the extreme 2010 drought, over 5 000 km^2 and during the El Niño in 2015/16, over 9 000 km^2 of forests burned (Silva Junior et al., 2019). Estimates focused on southern Amazonian forests suggest that 12% and 5% of forests burned during 2007 and 2010 (Paulo Monteiro Brando et al., 2014). All these extent of fire-affected forests, coupled with the increase in fragmentation (Silva Junior et al., 2021) is turning these forests into a different ecosystem state (Berenguer et al., 2014), with a long-term negative impact on its functioning (Silva et al., 2020).

Large-scale Amazon ecosystem shifts are nevertheless likely in the absence of climate mitigation, would severely impact local populations by radically altering regional ecology and climate, and could influence climate patterns around the world. Such changes have been hypothesized to potentially occur within a relatively short time frame of around 50 years (K. H. Cook & Vizy, 2008; Fonseca et al., 2019; Lyra et al., 2016), significantly reducing the strength of the Amazon system as a terrestrial natural carbon sink and also releasing considerable quantities of stored carbon to the atmosphere via decomposition, wildfire, and land-use changes.

3.2.2 Substantial evidence indicates that significant regions of the Amazon are at risk of dieback caused by fires and drought

Both observations as well as modeling studies have proposed a critical threshold for Amazon forest dieback, although ongoing debate continues to discuss flaws in this hypothesis and contest whether the region will respond to deforestation and climate change as a biome-wide tipping element. Uncertainties also complicate efforts to precisely determine the boundaries of regional critical thresholds. Nevertheless, the general threat of regional ecosystem shifts driven by wildfires, drought, and deforestation is well agreed upon within the research community.

Water represents a key variable for assessing the Amazon's future. The Amazon rainforest typically receives annual precipitation of approximately 2 200 mm; the western reaches of the region receive the highest amount of rainfall (3 000mm) thanks to the topographic influence of the Andes mountains (Salati & Vose, 1984), and these wettest portions of the rainforest are likely to survive even under drier conditions triggered by dieback. The lower limit of precipitation necessary to maintain a closed-canopy tropical rainforest sits at around 1 600 mm/yr (Hirota, Holmgren, Van Nes, & Scheffer, 2011). The dry season, defined as when the rainfall is lower than evapotranspiration, varies from zero to six months in western to southern Amazonia, respectively (Berenguer et al., 2021). During the dry season, the Amazon region experiences a water deficit that is partially replenished by recharge during the wet season. Consequently, the twin factors of dry season severity and overall rainfall play central roles in maintaining existing rainforest cover (Malhi et al., 2009).

Observations are already indicating that the regional water cycle is shifting towards conditions of increasing water stress. The southeastern Amazon basin currently experiences reduced rainfall relative to the rest of the basin, around 1 700 mm/yr - an effect that is attributed to higher intensity of land use (Salati & Vose, 1984). The frequency of drought events in the southern Amazon has also increased according to a study examining dry events from 1970-1999 (W. Li, Fu, Juárez, & Fernandes, 2008). Significant evidence suggests that the dry season has lengthened, particularly over the southern and southeastern Amazon (Dubreuil, Debortoli, Funatsu, Nédélec, & Durieux, 2012; R. Fu et al., 2013; Marengo, Nobre, Sampaio, Salazar, & Borma, 2011). Moreover, observations have shown an increasing number of months with a cumulative water deficit and a rainier wet season in the southern region (Anderson et al., 2018), and an increase in temperature of 1.6°C to 2.5°C from the west to south parts of the Amazon between the months of August and October over the last 40 years (Gatti et al., 2021). CMIP6 models demonstrate general agreement that regional rainfall will continue to decline over the 21st century under unmitigated climate scenarios (L A Parsons, 2020).

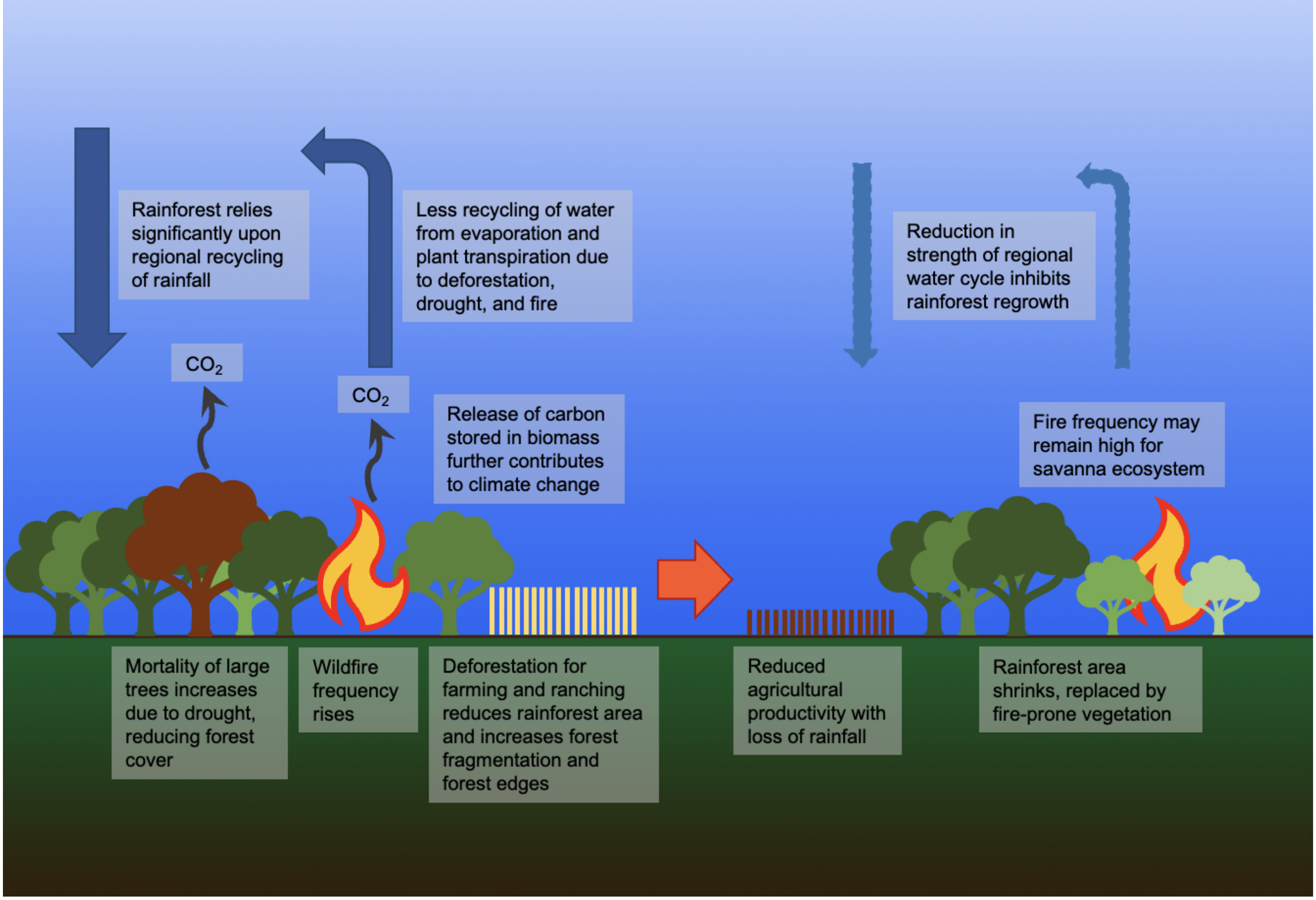

Figure 6: Schematic diagram of causes, feedbacks, and impacts associated with Amazon rainforest mortality.

At the same time, multiple factors exert important influences over Amazon ecosystem health (Figure 6). Increasing regional temperatures due to both local and global drivers (Jones, 1994; Victoria et al., 1998) elevate potential evaporation and reduce humidity, increasing quantities of dry fuel and heightening fire risk (Cochrane & Barber, 2009; Cochrane & Schulze, 1999; Pueyo et al., 2010). The seasonality and magnitude of rainfall may change due to shifts in the El-Nino Southern Oscillation (Aragão et al., 2018), although predictions remain highly uncertain. Some research has suggested that elevated CO_2 concentrations may reduce the rate of stomatal leaf opening in trees, lowering evapotranspiration and thereby further weakening recycled precipitation (Lammertsma et al., 2011; A. P. Walker et al., 2014).

The influence of these diverse factors complicates efforts to more precisely define the critical boundaries to Amazon rainforest stability. A temperature increase of 3-4°C (Carlos A. Nobre et al., 2016), deforestation of >40% (Carlos A. Nobre et al., 2016), or precipitation decrease of 30-40% (Lenton et al., 2008; Salazar & Nobre, 2010) each have been proposed as thresholds independently capable of initiating Amazon dieback according to models, but all three factors are currently occurring simultaneously with mutually-reinforcing effects (Staal, Flores, et al., 2020). Based on rainfall levels corresponding to current rainforest extent, modeling of projected hydrological changes under high-emissions scenarios suggest the potential for committed Amazon tipping behavior by end-of-century driven by rainfall shifts alone (Staal, Fetzer, et al., 2020). Lovejoy and Nobre have proposed a deforestation extent of 20-25% as the tipping point of savannization, suggesting that the current Amazon system already stands perilously close to a critical threshold (Lovejoy & Nobre, 2018), beyond which a strong self-sustaining water stress feedback will act to convert 50-70% (K. H. Cook & Vizy, 2008; Lyra et al., 2016) of the region to a savanna-type ecosystem. These thresholds have been defined based on an intact forest perspective, and currently, with the increased area of human-modified forests, either through fires of logging degradation which impacts its functioning, it could be expected that a lower threshold could be considered.

Substantial uncertainty characterizes modeled future precipitation changes for the region (Good et al., 2013; W. Li, Fu, & Dickinson, 2006; L A Parsons, 2020) and models' ability to represent subsurface hydrological processes remains limited (Brodribb, Powers, Cochard, & Choat, 2020; National Research Council, 2013). In particular, some researchers have asserted that increases in wildfires as a result of drying will play a more immediate and serious role in driving forest loss (P.M Brando et al., 2019). Field studies have indicated that while prolonged drought does eventually raise mortality rates for rainforest trees, this only occurred after the third consecutive year of a particularly intense simulated drought (Nepstad, Tohver, Ray, Moutinho, & Cardinot, 2007). In contrast, a forest fire during a drought period can kill half the adult trees in the affected area within a much shorter period (Balch et al., 2015; Paulo Monteiro Brando et al., 2014; Nepstad et al., 1999). Recovery of fire-affected forest can require many decades, resulting in strong net CO_2 emissions from long-term reductions in carbon storage (Silva et al., 2020). Tree mortality and reduced canopy cover caused by fire can in turn permit invasive, more fire-prone grasses to invade the forest floor, increasing the region's susceptibility to fire and driving conversion to grassland that primarily results from dryness and fire, not drought-induced water stress (Balch et al., 2015). A recent remote sensing study examining changes in aboveground biomass across Amazonia over the last decade supports the hypothesis that forest degradation may already drive more biomass loss than deforestation (Qin et al., 2021). A newly-published aerial survey of carbon emissions over Amazonia concludes that southeastern Amazonia has transitioned to a net source of atmospheric carbon due to the effects of deforestation, fire, and shifts in regional climate (Gatti et al., 2021).

Recent advances in soil-plant hydraulic modelling have also highlighted higher general potential vulnerability of forest ecosystems to drought than previously suggested by traditional ecological niche-based models (Brodribb, Cochard, &

Dominguez, 2019; Brodribb et al., 2020; Martin-StPaul, Delzon, & Cochard, 2017). While knowledge gaps remain regarding plant physiological modelling and the ability of trees to adapt or acclimate to long-term increases in water stress, such newer approaches are producing more rapid increases in tree mortality in response to climate-induced drought (Brodribb et al., 2020).

3.2.3 Loss of Amazon rainforest could occur over decades to a century and cause important climate, human, and ecosystem impacts

According to the rainfall-driven dieback theory, after crossing a critical threshold, affected portions of the Amazon would undergo savannization over a timeframe of likely somewhere between several decades to a century (Lenton et al., 2008; Lovejoy & Nobre, 2018; Nepstad, Stickler, Soares-Filho, & Merry, 2008; Carlos A. Nobre, Sellers, & Shukla, 1991; Carlos Afonso Nobre & Borma, 2009). Collapse begins with die-off of the largest trees as soil moisture becomes increasingly depleted (Ivanov et al., 2012; Nepstad et al., 2008), progressively thinning the forest canopy. This transition would likely be accompanied by intensification of wildfires as dead vegetation and drier conditions contribute towards higher fire risk (Paulo Monteiro Brando et al., 2014; Cochrane & Barber, 2009; Cochrane & Schulze, 1999). A modeling study projects significant future increases in land area with a fire relative probability of 0.3 or greater under RCP4.5 emissions scenarios (+21.3% increase in area with FRP>0.3), with substantially greater fire risk (+113.5% increase in area with FRP>0.3) for a high-deforestation RCP8.5 scenario (Fonseca et al., 2019). This drought-fire feedback might in fact independently drive ecosystem shifts, dominating over impacts from water stress and resulting in a pattern of forest loss driven by more local factors as opposed to a biome-wide threshold (Balch et al., 2015; P.M Brando et al., 2019; Paulo Monteiro Brando et al., 2014).

Degradation and areal reduction of the Amazon rainforest might even still occur under strong climate mitigation scenarios, with significant biomass losses possible at 1.5-2°C of warming (O. Hoegh-Guldberg et al., 2018). Deforestation can also drive fragmentation of forest landscapes, leading to forest degradation and significant carbon losses through edge effects (Silva Junior et al., 2020). However, forest management policy could act as a strong lever upon the Amazon’s ecosystem health, with a halt to new deforestation potentially reducing burned area by 2050 by 30% and accompanying greenhouse gas emissions by 56% (P.M Brando et al., 2019). Fire suppression also represents a potentially powerful tool to reduce rainforest loss by preventing positive feedbacks between fire and savannization. As the Amazon rainforest is naturally warm, forest litter decomposes rapidly and does not accumulate as it does in some temperate forests, meaning that aggressive fire suppression carries no risk of intensifying future fires.

Once savannization or conversion to a more seasonal forest has occurred, significant hysteresis may inhibit the return of rainforest vegetation even over very long timescales, thanks to potentially more frequent fires and drier climate acting as conditions upon the vegetation's ability to to regenerate dense tropical forest and large tree species. Paleoclimate evidence confirms that regrowth of rainforest cover often lagged significantly behind the return of wet conditions to South America during past periods of the Holocene (Ledru, Salgado-Labouriau, & Lorscheitter, 1998).

Loss of large portions of the Amazon rainforest would carry significant implications for the strength of the Amazon carbon sink as well as the fate of the organic carbon within Amazon vegetative and soil biomass. Rainforest degradation could likely shift the Amazon region from a carbon sink of up to 0.6 Pg/yr (Malhi et al., 2006) to a potentially strong carbon source (Carlos Afonso Nobre & Borma, 2009). Modeling of an Amazon rainforest dieback scenario under moderate emissions (713 ppm in 2100) produced an additional 0.3°C of warming globally as a result of rainforest loss (R. Betts, Sanderson, & Woodward, 2008).

Precise estimates of the carbon impact of Amazon dieback are however subject to variability due to uncertainty regarding the potential extent of forest loss, the current size of the Amazon forest carbon pool, and potential timescales of change. A recent modelling analysis using CMIP6 rainfall projections under a high-end RCP8.5 emissions scenario found a large potential reduction in stable Amazon forest area of up to 3.27 million km^2 due to hydrological changes (Staal, Fetzer, et al., 2020). Under a RCP8.5 emissions pathway with continued deforestation, carbon dioxide emissions from intensifying fires were estimated to sum to 4.6 Gt by 2050 (P.M Brando et al., 2019). However, carbon release as a result of forest fire can eventually be partially offset by recovery of vegetation within the burned areas (Silva et al., 2020). Carbon emissions from losses in forest area may also be potentially compensated for if the positive response of rainforest vegetation to increased carbon dioxide levels is large (strong CO_2 fertilization). As the effect of CO_2 fertilization upon plant growth remains highly uncertain, this complicates efforts to estimate the overall carbon impact of the world's tropical forests as they respond to climate change (Cox et al., 2013; Huntingford et al., 2013). However, recent observational evidence strongly suggests that the CO_2 fertilization effect is nearing saturation in African tropical forests as well as the Amazon on faster-than-predicted timescales (Hubau et al., 2020).

Amazon rainforest loss possesses acute regional ramifications. Degradation of the rainforest represents a serious threat to the livelihoods and lifestyles of the traditional Amazonian population. Total rainfall throughout the Amazon basin may generally fall following dieback and savannization, placing regional agriculture at risk (Arvor, Dubreuil, Ronchail, Simões, & Funatsu, 2014; Leite-Filho et al., 2021; Oliveira, Costa, Soares-Filho, & Coe, 2013). A study has found that for 40% deforestation, hydroelectric power generation from major dams on the Xingu River is substantially reduced thanks to falling river runoff (Stickler et al., 2013). That said, the complexity of rainfall dynamics means that impacts depend strongly on the spatial pattern of forest loss (D. Lawrence & Vandecar, 2015). Increased wildfire frequency and severity will additionally put regional communities at risk and create air pollution crises. Finally, dieback of the Amazon rainforest will represent a major threat to the biodiversity of this region (Esquivel-Muelbert et al., 2017; Gomes, Vieira, Salomão, & ter Steege, 2019).

Among the tipping elements covered throughout this review, the Amazon forest dieback scenario stands among the more imminent, likely, and fast-acting of the climate tipping elements popularly discussed within the climate community. Substantial scientific evidence backs the assessment that the Amazon rainforest faces a serious threat within the 21st century, with deforestation and wildfire playing a particularly critical role in driving the ecosystem towards a sudden transition. Forest management policy will prove key to determining the future of the Amazon, as evidenced by the success of the Brazilian government in at least temporarily reducing the rate of deforestation relative to the high pace of loss seen in the early 2000s (Carlos A. Nobre et al., 2016). Under present pressure, however, the Amazon rainforest will indeed face serious threats this century like many other earth systems, with significant regional and global consequences accompanying its degradation.

3.3 Loss of summer Arctic sea ice

3.3.1 Background

The shrinking of Arctic sea ice area over the late 20th and early 21st centuries has been clearly documented by observations of decreasing sea ice extent in all months and all regions of the Arctic (Julienne Stroeve & Notz, 2018), with strong reductions in area during summer within both the Pacific and Eurasian sectors (Årthun, Onarheim, Dörr, & Eldevik, 2021; Fetterer, Knowles, Meier, Savoie, & Windnagel, 2017; Onarheim, Eldevik, Smedsrud, & Stroeve, 2018; J. C. Stroeve, Serreze, et al., 2012). While summer ice loss has been dramatic, departures from average conditions have been largest in the shoulder seasons (Julienne Stroeve & Notz, 2018). From 1997 to 2007, the Arctic ice cap has shrunk by an area of 1.5 million km2 (Nghiem et al., 2007). Reduction in sea ice extent over the observational record has at times outpaced older IPCC model projections (Julienne Stroeve, Holland, Meier, Scambos, & Serreze, 2007). While the spatial extent of more marginal sea ice—areas of the Arctic Ocean with partial ice cover—has remained relatively constant over the past 40 years, its proportion as a fraction of total sea ice is increasing (Rolph, Feltham, & Schröder, 2020).

The characteristics of Arctic sea ice have also shifted, with multi-year sea ice at least five years old falling from 30% of Arctic ice to just 2% between 1984 and 2019 while first-year sea ice has increased from 40% to 60-70% of Arctic sea ice area (Julienne Stroeve & Notz, 2018). At the same time, mean winter multi-year ice thickness has declined from 3.6 m to 1.9 m over a thirty-year period from ~1980-2010 (Kwok & Rothrock, 2009; Wadhams, 2012).

The decline in Arctic sea ice is statistically attributable to a strong anthropogenic forcing from greenhouse gases (Notz & Marotzke, 2012), and also reflects profound changes in the timing of melt onset and autumn freeze-up (Notz & Stroeve, 2016, 2018). In addition to regional warming, several positive feedbacks are responsible for the rapid pace of shrinking sea ice extent. The well-known ice-albedo feedback, in which the melt-driven substitution of highly-reflective sea ice for highly-absorptive dark open ocean waters results in increased surface inputs of solar energy, contributes to further ice loss via ocean and lower atmospheric warming. Sea ice loss is marked by an increasing transition from multi-year to seasonal sea ice, which is thinner and therefore more vulnerable to melt (Haine & Martin, 2017). Reduced cooling from lowered summertime cloud cover may also be exerting a potential effect (Kay, L'Ecuyer, Gettelman, Stephens, & O'Dell, 2008).

Other feedback processes such as earlier melt onset have enhanced the ice-albedo feedback (J. C. Stroeve, Serreze, et al., 2012), increasing solar heat inputs across 85% of the Arctic region between 1979-2007 (Perovich et al., 2007) and driving a fast-paced increase in regional ocean temperatures of around 0.5C per decade (Timmermans, Ladd, & Wood, 2017) that further melts summer ice cover. With larger areas of open water at the end of summer and increased ocean mixed-layer temperatures, autumn freeze-up is delayed as it takes time for the ocean to release the heat gained over summer back to the atmosphere. Thus, freeze-up trends largely drive the expanding open water period e.g. (J. Stroeve, Barrett, Serreze, & Schweiger, 2014; Julienne Stroeve & Notz, 2018).

Some negative or stabilizing feedbacks do exist, moderating the pace of sea ice loss. Thinner first-year sea ice grows more rapidly during periods of freezing, while large areas that are ice-free or dominated by first-year ice lose heat to the atmosphere faster during winter, allowing thin ice cover to regrow quickly over large areas (Ian Eisenman, 2012; Notz, 2009; Sturm & Massom, 2016; T. J. W. Wagner & Eisenman, 2015). While the later onset of freezing is a contributing factor to reduced sea ice extent, this shift in timing also limits snow accumulation that would otherwise insulate the ice and inhibit growth (Sturm & Massom, 2016). These effects mean that years with very low summer ice coverage could enhance recovery of ice area the following winter, leading to a rebound in ice extent (Bitz & Roe, 2004). Overall, however, such negative feedbacks are only marginally mitigating the rapid loss of Arctic sea ice.

3.3.2 Rapid decline of Arctic sea ice extent risks episodic ice-free summers before mid-century

The current research literature points towards a highly linear, predictable response of summer sea ice extent in response to greenhouse gas emissions, despite large interannual variability in weather patterns, rather than an abrupt transition to seasonally ice-free conditions consistent with tipping behavior (I. Eisenman & Wettlaufer, 2009; Notz & Stroeve, 2016; Julienne Stroeve & Notz, 2018; Tietsche, Notz, Jungclaus, & Marotzke, 2011; Winton, 2011). The linear relationship between sea ice area and cumulative greenhouse gases in the atmosphere reveals the Arctic is already in the process of transitioning towards ice-free summer conditions, with the first ice-free summer potentially occurring before mid-century (Notz & Stroeve, 2018; SIMIP Community, 2020). The complete loss of summer Arctic sea ice would represent an important shift for regional climate and ecology, triggering significant consequences for regional warming and for Arctic biodiversity. For high levels of warming, a possibility also emerges of a more rapid transition towards an ice-free Arctic in winter, once ice-free summer conditions become more common (Bathiany, Notz, Mauritsen, Raedel, & Brovkin, 2016; I. Eisenman & Wettlaufer, 2009).

Unless greenhouse gas emissions are reduced, it is likely the Arctic will lose its summer ice cover within current lifetimes. While CMIP5 model projections predicted a high likelihood of ice-free summers in the second half of the 21st century without strong climate mitigation (Massonnet et al., 2012; J. C. Stroeve, Kattsov, et al., 2012), CMIP6 models suggest the first ice-free Septembers will happen before 2050 (SIMIP Community, 2020). Some uncertainty does exist regarding the exact timing when ice-free summers might be expected to manifest, based simply on the large range of internal variability, with estimates for the approximate date of onset of occasional seasonally ice-free years varying by +/-20 years (Meredith et al., 2019). Neverthless, the strong relationship between ice loss and greenhouse forcing coupled with current emissions rates indicates a reasonable likelihood for the first ice-free summer to occur before mid-century (Julienne Stroeve & Notz, 2018). Modeling studies suggest that nothing short of extremely aggressive climate mitigation is likely to reduce the likelihood of future summer sea ice loss; only for warming of no more than 1.5°C is sea ice generally still present in summer by end-of-century (Jahn, 2018; Notz & Stroeve, 2018). For the majority of new CMIP6 models, the Arctic Ocean experiences its first ice-free September before 2050 at 1.5°C pathways (SIMIP Community, 2020). However, CMIP6 models continue to exhibit a very wide range of spread and performance, reflecting internal variability.

While loss of summer Arctic sea ice represents a high-likelihood outcome under current rates of warming, a totally ice-free Arctic year-round remains a relatively much less probable scenario. Loss of winter Arctic sea ice is assessed to be possible, but only under worst-case emissions scenarios (RCP8.5) (Bathiany et al., 2016; Winton, 2006). One modeling analysis found that ice-free winter conditions required around 13°C mean annual warming at the north pole (Winton, 2006), and is thus consequently unlikely to occur this century. Loss of winter sea ice may however proceed more abruptly than the long-term decline in summer sea ice extent to date, potentially occurring just a few years after the ocean becomes too warm to form ice in winter (Bathiany et al., 2016). The homogenously thin state of winter sea ice under conditions of frequent ice-free summers further contributes to its rapid loss. At the same time, loss of winter sea ice is thought to be reversible based on modeling results, merely requiring that winter sea temperatures fall back below freezing thresholds (Armour, Eisenman, Blanchard-Wrigglesworth, McCusker, & Bitz, 2011; Bathiany et al., 2016; C. Li, Notz, Tietsche, & Marotzke, 2013; Ridley, Lowe, & Hewitt, 2012).

Areas of uncertainty do remain when attempting to model future Arctic sea ice area. Cloud feedbacks that are important for determining summer sea ice dynamics still remain challenging even for state-of-art models (Tietsche et al., 2011). Poor representation of the varying distribution of sea ice thicknesses over the Arctic region is another weakness of current models

(Holland, Bailey, & Vavrus, 2011; J. Stroeve et al., 2014), likely due to varying ability to represent atmospheric circulation's influence on ice thickness (Bitz, Fyfe, & Flato, 2002; Kwok & Rothrock, 2009; J. Stroeve et al., 2014). The influence of seasonal atmospheric processes upon sea ice dynamics remains incompletely understood and an active area of research (Topal et al., 2020). Nevertheless, such potential sources of error do not meaningfully affect the conclusion that ice-free summers across the Arctic Ocean will likely begin to occur starting before mid-century.

3.3.3 Loss of summer sea ice accelerates regional warming with global implications

With much of the region under permanent or extended daylight hours during summer months, ice-free conditions lead to a high potential for accelerated regional warming via the aforementioned ice-albedo feedback. Reductions in Arctic sea ice extent are implicated in enhanced Arctic warming that cannot be alternatively explained by cloud cover changes, internal climate variability, or variability in atmospheric and oceanic circulation (Screen, Deser, Simmonds, & Tomas, 2014; Screen & Simmonds, 2010; Serreze, Barrett, Stroeve, Kindig, & Holland, 2009). Modeling analysis also indicates a strong sensitivity of Arctic temperatures to sea ice loss (Screen et al., 2014). This effect has consequently played a strong role in driving rates of regional warming at more than double the global average rate, and is expected to continue to contribute to regional amplification of warming (Meredith et al., 2019). The scientific community is continuing to investigate the potential impact of Arctic sea ice decline upon weather and climate patterns in the Northern Hemisphere (Barnes & Screen, 2015; M. Kretschmer, Zappa, & Shepherd, 2020).

This ice-albedo feedback additionally influences global climate at large. A modeling analysis examining a future Arctic with seasonally absent sea ice in summer found that an ice-free summer Arctic accelerated the rate of global warming by ~8%, increasing radiative forcing by 0.3 W/m^2, a warming contribution equal to that of atmospheric halocarbons (Hudson, 2011). A more recent analysis obtained a figure of 0.49 W/m^2 for a somewhat higher summer ice loss scenario in which summer sea ice remains low for a five-month period (Nico Wunderling, Willeit, Donges, & Winkelmann, 2020).

The regional warming impact of significant reductions in Arctic sea ice extent also presents an indirect threat to global sea-level rise by promoting additional melt from the Greenland Ice-sheet, an interaction identified as important over the geologic past in paleoclimate models (Koenig, DeConto, & Pollard, 2014).

Finally, the loss of summer Arctic sea ice carries significant ecological implications. Wildlife dependent on sea ice for shelter or survival may be severely impacted, and the transition to ice-free summer conditions will also likely cause substantial shifts to phytoplankton community structure, driving transitions in regional marine ecology (Meredith et al., 2019). These impacts to wildlife and fishing will likely present economic, social, and cultural challenges for human communities across the Arctic.

A year-round ice-free Arctic remains a scenario not currently believed to be plausible until 2100 even under worst-case warming scenarios (Bathiany et al., 2016; Lenton, 2012; Winton, 2006). Furthermore, loss of Arctic sea ice can be reversed on shorter timescales than most of the other climate mechanisms discussed in this review, as sea ice extent can recover within decades if initial warming is reversed (Notz, 2009; Ridley et al., 2012). At present however, the Arctic sea ice system confronts an accelerated rate of warming, a rapid pace of sea ice decline, and an increasing likelihood of ice-free summer conditions occurring within a couple decades.

4 Tipping element cascade leading to irreversible "Hothouse Earth" warming

4.1 Background

In 2018, an article by (Steffen et al., 2018) proposed a "Hothouse Earth" scenario, in which modern anthropogenic warming could trigger strong positive climate feedbacks leading to significant temperature increases well beyond those expected from human greenhouse gas forcing alone, potentially altering the long-term climate trajectory of the Earth itself, removing planetary temperature variations from the glacial-interglacial cycle and placing the biosphere in a newly-created warmer equilibrium that could persist for up to hundreds of thousands of years.

Historically, the Earth has transitioned between numerous climate states, from the very warm Paleocene-Eocene Thermal Maximum 56 million years ago (McInerney & Wing, 2011) to the glacial and interglacial cycles of the Quaternary Period (2.6 million years ago to present). Within the latter period, which encompasses the entirety of recorded human history, shifts between glacial and interglacial periods have been primarily driven by slow changes in the Earth's orbit on timescales of approximately 100 000 years. Anthropogenic climate change, however, is driving changes in the earth system at a rapid pace that the planet has rarely experienced apart from cataclysmic events such as asteroid impacts (Zeebe, Ridgwell, & Zachos, 2016).

Indeed, a number of studies argue that current human greenhouse gas emissions and associated warming will likely prove sufficient to delay the onset of the next ice age by 50 000 years or more (Berger & Loutre, 2002; Ganopolski, Winkelmann, & Schellnhuber, 2016). Cumulative emissions of over 1 000 Gt C might be sufficient to produce a prolonged postponement of glaciation, which would include all RCPs with emissions exceeding an RCP2.6 scenario (Ganopolski et al., 2016). Given the high importance of atmospheric greenhouse gas levels, alongside cycles in the Earth's long-term orbit, for driving glacial-interglacial shifts, climate change will also likely alter the dynamics of the transitions between glacial and interglacial states, potentially amplifying or accelerating the response of global climate to orbital forcings, as reviewed in (Masson-Delmotte et al., 2013).

Paleoclimate evidence suggests that past climate states of the Earth may have shared similarities with a warmer future world impacted by anthropogenic climate change, including the Early Eocene (50 million years ago) and the Mid-Pliocene (3.3 - 3 million years ago) (K. D. Burke et al., 2018). Such periods were marked by notably warmer mean surface temperatures (Early Eocene: 13±2.6°C warmer than present day; Mid-Pliocene 1.8-3.6°C warmer than present day), reduced (Mid-Pliocene) or absent (Early Eocene) ice-sheets, elevated sea-levels, and higher CO_2 concentrations (Early Eocene: 1 400 ppm; Mid-Pliocene: 400 ppm). Such paleoclimate analogs for a warmer earth such as the Eocene may also have been characterized by a greatly-reduced extent of subtropical clouds (Schneider et al., 2019). The Mid-Miocene (15.5 to 17 million years ago, 2-4°C warmer, 300-500 ppm CO_2) (Greenop, Foster, Wilson, & Lear, 2014; Kominz et al., 2008) also holds some potential parallels with a warmer Earth. The range of conditions denoted by these past eras support the idea that a similar hotter future climate could remain stable over geologic time.

In describing the path towards a "Hothouse Earth" equilibrium, (Steffen et al., 2018) suggest that failure to meet a 2°C warming goal may risk triggering one or more climate tipping elements with relatively low critical warming thresholds. Additional warming brought on by these more sensitive tipping elements would in turn activate a number of other tipping elements in a self-reinforcing cascade, potentially pushing the climate system several degrees hotter in response to even a modest failure to meet a 2°C warming threshold. Due to the extremely long response time of natural carbon sinks that would eventually draw down atmospheric CO_2 released through human activity and the ensuing "tipping point cascade", the authors express concern that such a Hothouse Earth climate could last for hundreds of millennia - a timescale comparable to the length of glacial - interglacial cycles. A newer Nature Comment presented an updated risk assessment of a tipping element cascade with the argument that warming of 1-2°C globally already represents an unacceptably high risk of triggering major tipping elements (Lenton et al., 2019). A recent study using a conceptual network simple modeling approach similarly stressed the potential for tipping elements to interact, which could shift true critical thresholds within their potential ranges (N Wunderling et al., 2021). All three articles nevertheless emphasize substantial disagreement and uncertainties surrounding key assumptions regarding tipping elements and their interactions, stressing the need for more quantitative and modeling analysis to substantiate the risks of significant additional climate change as well as the long timescales required for some of the climate mechanisms involved to act.

In the following subsections, we discuss the likelihood of significant additional climate change driven by tipping elements with the aid of a simple climate model (FaIR v1.6.3) (Millar et al., 2017; Christopher J. Smith et al., 2017). Drawing upon our literature synthesis throughout this review, we suggest that is unclear whether near-term tipping elements possess the potential to drive large (>1°C) additional warming over the next couple of centuries (2100-2300), while long-term tipping elements act sufficiently slowly that their impacts lower the risk of cascading effects and can be potentially mitigated by societal action. Other tipping elements, such as coral reef loss or ice-sheet retreat will have minimal effects on global mean surface temperatures. Such considerations are important to include when assessing and discussing tipping element cascade theory. Nevertheless, knowledge gaps associated with many tipping elements and their potential interactions with one another and global carbon cycle feedbacks highlight a priority need for additional research.

4.2 Imminent tipping elements

Evaluation of the Hothouse Earth hypothesis involves considerable complexity, as rather than representing its own independent mechanism, the Hothouse Earth pathway effectively consists of numerous individual tipping elements and teleconnections between them.

Discussion of the Hothouse Earth hypothesis thus involves three questions: 1) what warming thresholds are tipping elements activated at, 2) are tipping elements with lower thresholds sufficient to drive warming that activates tipping elements with higher thresholds, and 3) what is the cumulative effect of all activated tipping elements and resulting climate feedbacks?

Earth system elements identified by (Steffen et al., 2018) that this review indicates are at high risk of committed change in response to warming this century include loss of Arctic summer sea ice, loss of portions of the Greenland Ice-sheet, loss of portions of the West Antarctic Ice-sheet, Amazon rainforest dieback, boreal forest dieback, some permafrost carbon release, potential destabilization of the AMOC, and coral reef loss (Figure 7). Significant ice mass loss from mountain glaciers this century is also virtually certain (IPCC, 2019).

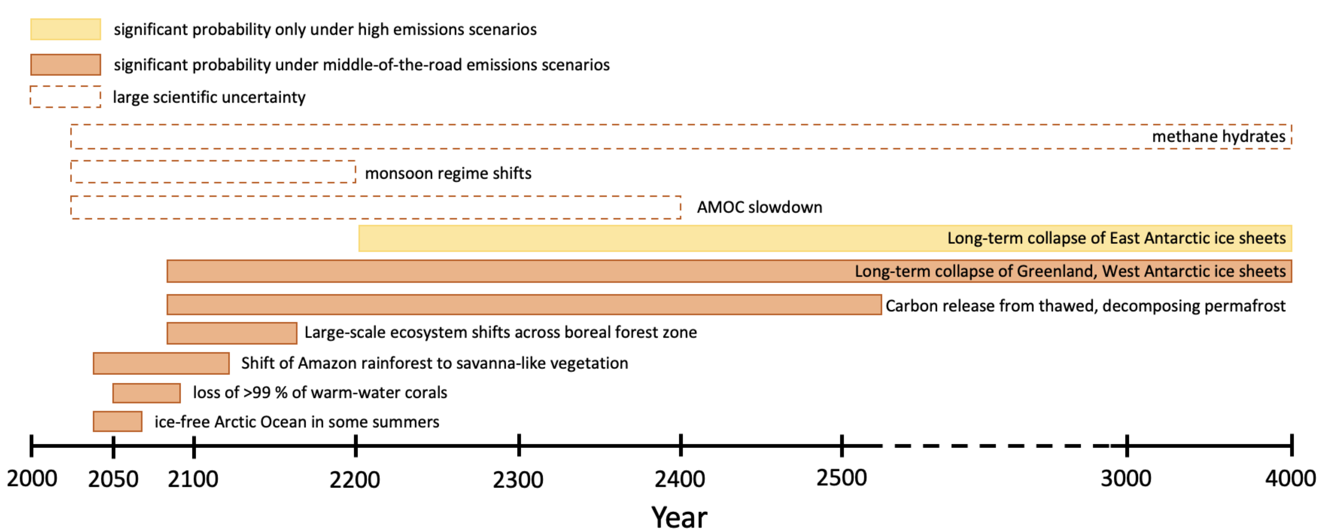

Figure 7: Illustration of the approximate time frames over which earth system elements, including tipping elements, might be expected to respond to climate forcings.

While irreversible ice mass loss from the Greenland Ice-sheet and West Antarctic Ice-sheet (Section 2.3) and weakening of the Atlantic Meridional Overturning Circulation AMOC (Section 2.1) are likely to be initiated by current rates of warming over the next 100 years, prevailing scientific opinion holds that even under worst-case warming scenarios these mechanisms will likely take multiple centuries before strong climate impacts manifest (P. Bakker et al., 2016; Peter U. Clark et al., 2016; N. R. Golledge et al., 2015; Huybrechts & De Wolde, 1999). Consequently, these three longer-term tipping elements are unlikely to help initiate a tipping element cascade by driving significant climate change in the near-term. We will instead discuss these mechanisms in the next section covering long-term tipping elements.

Of the major tipping elements likely to be triggered this century under current trends, Amazon rainforest dieback (see Section 3.2) represents a large uncertain factor, with high potential to contribute significantly towards climate change on a relatively fast timeframe. This stems from the considerable range in estimates of the size of the Amazon forest carbon pool, the strength of the current Amazon carbon sink (0.4 - 0.6 Gt C/yr) (Malhi et al., 2006; Phillips et al., 2009), and the potential magnitude and timescales of conversion of rainforest to a different ecosystem (50-70%, 50-100 years) (K. H. Cook & Vizy, 2008; Lyra et al., 2016). A study simulating the climate impact of Amazon rainforest dieback under a moderate emissions scenario (IS92a, with CO_2 levels reaching 713 ppm in 2100) estimated that rainforest loss causes around an additional 0.3°C of warming (R. Betts et al., 2008). Some researchers have proposed that the Amazon forest could reach a tipping threshold within coming decades due to the combined stresses of deforestation and warming (Lovejoy & Nobre, 2019).

Boreal forest dieback (Section 2.5) also represents a potentially imminent and impactful climate tipping element. The circumpolar boreal forests store large quantities of carbon and are currently being subjected to increasing stress from fires (De Groot et al., 2013; Flannigan et al., 2009; Shvidenko et al., 2011), pests (Kurz et al., 2008), and rising temperatures (Screen & Simmonds, 2010), exhibiting elevated rates of tree mortality over the observational record (Allen et al., 2010). Modeling studies suggest a credible potential for large-scale biome shifts across major boreal regions in response to high-emissions climate scenarios over the next century (Foster et al., 2019; Mann et al., 2012; Shuman et al., 2015; Wu et al., 2017). The sign, speed, and magnitude of the climate impact of boreal forest changes considering carbon fluxes and albedo changes remain uncertain, although current research cannot eliminate the possibility for a sizable net warming contribution to global climate.

While gradual Arctic permafrost thaw (Section 2.4) predominantly represents a slow process occurring over centuries, gradual thaw could also contribute significant additional carbon emissions over the near-term (92 Gt C by 2100 under worst-case scenarios) (Meredith et al., 2019). Abrupt permafrost thaw processes acting over much more immediate timescales could emit up to ~18 Gt C by 2100 including a significant quantity of methane (M. Turetsky et al., 2020; M. R. Turetsky et al., 2019). Over this century, emissions from abrupt thaw could contribute approximately 6 771 Mt CH_4 (Mt C) and 10.95 Gt CO_2 (Gt C) under the worst-case RCP8.5 scenario (M. Turetsky et al., 2020), compared to current annual human emissions of 335 Mt C/yr of methane and 10 Gt C/yr of CO_2. Combined, gradual and abrupt permafrost carbon emissions by 2100 will likely meaningfully reduce remaining carbon budgets for current climate targets, but are far from sufficient to independently drive substantial additional warming.

For Arctic sea ice (Section 3.3), a realistic modeling scenario for future Arctic sea ice loss, in which sea ice extent is considerably reduced with a completely ice-free period recurring for approximately one month every summer yields a change in radiative forcing of 0.3 W/m^2 – climatically significant, but only on a level similar to that of current anthropogenic forcing from atmospheric halocarbons (Hudson, 2011).

The remaining other imminent impacts – coral reef collapse (Section 3.1) and loss of mountain glaciers – will likely yield minimal to negligible additional warming or greenhouse forcing. From a climate perspective, it stands to reason that only tipping elements resulting in significant changes to the Earth's energy balance - through greenhouse gas emissions or albedo changes - increase the risk of a tipping point cascade. Neither coral reef collapse nor mountain glacier melt are anticipated to meaningfully alter greenhouse gas fluxes or planetary-scale albedo. Meanwhile, the likelihood of monsoon regime shifts remains a subject of active debate (Section 2.6).

To estimate the potential cumulative impact of these near-term tipping elements and Arctic sea ice loss on future climate projections through 2300, we leveraged the FaIR simple climate model (Millar et al., 2017; Christopher J. Smith et al., 2017) v1.6.3, adding further CO_2 emissions, CH_4 emissions, and radiative forcing changes produced from the response of tipping elements and Arctic sea ice to climate change under the SSP2-4.5 and SSP5-8.5 scenarios, as detailed in (Table 5). For our "base" scenarios, we generated anthropogenic emissions inputs for FaIR using the MESSAGE-GLOBIOM SSP2-4.5 and REMIND-MAGPIE SSP5-8.5 annual mean emissions specified in the RCMIP protocol (https://www.rcmip.org/). Volcanic and solar forcings were taken from the same RCMIP scenarios. Emissions rates and forcings were linearly interpolated for individual years. We utilized a natural emissions time series updated for the CMIP6 by Chris Smith following the methodology of (C J Smith et al., 2018).

	SSP2-4.5		**SSP5-8.5**		**Reference(s) as basis for assumption**
	CO_2 emissions/yr, Gt C	**CH_4 emissions/yr, Mt CH_4**	**CO_2 emissions/yr, Gt C**	**CH_4 emissions/yr, Mt CH_4**	
Gradual permafrost thaw	No net C fluxes (permafrost emissions compensated by	No net C fluxes (permafrost emissions compensated by	2010-2100: 0.55 (0.55 to 0.55) 2100-2200: 1.97 (0.50 to 3.41)	2010-2100: 5.19 (2.22 to 8.89) 2100-2200: 37.33 (4.67 to 116.67)	(McGuire et al., 2018; Schneider von Deimling et al., 2012)

	vegetation uptake)	vegetation uptake)	2200-2300: 0.98 (0.99 to 1.92)	2200-2300: 30.67 (14.67 to 104.00)	
Abrupt permafrost thaw	2000-2100: 0.032 (+/- 26%) 2100-2300: 0.083 (+/-36%)	2000-2100: 41.87 (+/- 26%) 2100-2300: 52.40 (+/-36%)	2000-2100: 0.01 to 0.18 (+/- 26%) 2100-2300: 0.21 to 0.34 (+/- 32%) Time-varying flux derived from Supp. Table 6 of (M. Turetsky et al., 2020).	2000-2100: 51.47 to 65.92 (+/- 26%) 2100-2300: 70.78 to 79.09 (+/- 32%) Time-varying flux derived from Supp. Table 6 of (M. Turetsky et al., 2020).	(M. Turetsky et al., 2020)
Methane hydrate destabilization	None, assume all C fluxes occur as methane	2000-2100: 2.37 (0 to 4.73) 2100-2200: 1.41 (0 to 2.82) 2200-2300: 1.23 (0 to 2.45)	None, assume all C fluxes occur as methane	2000-2100: 5.91 (2.37 to 9.46) 2100-2200: 3.52 (1.41 to 5.64) 2200-2300: 3.06 (1.23 to 4.9)	(K. Kretschmer et al., 2015)
Amazon forest dieback	2040-2100: 0.66 (+/- 15%) 2100-2150: 0.30 (+/- 15%)	None assumed.	2040-2140: 0.70 (+/- 15%) 2140-2200: 0.25 (+/- 15%)	None assumed.	Own assumptions, informed by (Aragão et al., 2018; Fonseca et al., 2019; Silva et al., 2020; Silva Junior et al., 2019, 2020)

	SSP2-4.5	**SSP5-8.5**	**Reference(s) as basis for assumption**
Arctic sea ice	Added radiative forcing climbs linearly from zero to 0.3 W/m^2 (+/-15%) by 2050, then to 0.5 W/m^2 (+/-15%) by 2100, remaining constant thereafter.	Added radiative forcing climbs linearly from zero to 0.3 W/m^2 (+/-15%) by 2030, then to 0.6 W/m^2 (+/-15%) by 2100, then to 0.7 W/m^2 (+/-15%) by 2120, remaining constant thereafter.	(Hudson, 2011; Pistone, Eisenman, & Ramanathan, 2019; Nico Wunderling et al., 2020)

Table 5: Additional CO_2 emissions, CH_4 emissions, and radiative forcing incorporated into SSP2-4.5 and SSP5-8.5 emissions pathways to simulate added climate impact of tipping elements and Arctic sea ice loss based on available estimates from literature. Values and percentages in parentheses indicate those selected under "high" and "low" scenarios.

For both the SSP2-4.5 and SSP8.5 pathways, we considered a "high", "middle", and "low" scenario, varying ECS and TCR accordingly and considering high, middle-ground, and low carbon fluxes and radiative forcing changes from tipping elements through 2300. For ECS, we chose low/median/high values of 2.3°K/3.1°K/4.7°K based on (Sherwood et al., 2020). For TCR, we selected a median value of 1.8°K based on (Sherwood et al., 2020) and low and high values of 1°K and 2.5°K based on the IPCC AR5 likely range for TCR.

Permafrost CO_2 and methane emission rates are derived from estimates of cumulative carbon release through 2300 for gradual permafrost thaw processes (McGuire et al., 2018), and for abrupt thaw processes (M. Turetsky et al., 2020) under RCP4.5 and RCP8.5, linearly interpolating C fluxes between dates for which net changes in ecosystem carbon stocks are reported. For emissions from gradual thaw, we assume 0.7% (0.3-1.2%) / 1.4% (0.7-2.5%) / 2.3% (1.1-3.9%) of carbon is released as CH_4 from 2010-2100 / 2100-2200 / 2200-2300, based on (Schneider von Deimling et al., 2012). We used described results for the SiBCASA, CLM4.5, and UVic RCP8.5 model runs in (McGuire et al., 2018) as the middle, low, and high cases respectively for carbon release from gradual thaw under SSP5-8.5 warming. For SSP2-4.5, we assume that carbon fluxes from permafrost degradation are balanced by vegetation effects (McGuire et al., 2018). We based low and high scenarios for CO_2 and CH_4 fluxes from abrupt thaw processes upon the one standard deviation ranges for RCP4.5 and RCP8.5 model results for the years 2100 and 2300 (M. Turetsky et al., 2020).

Methane hydrate emissions utilize estimates for cumulative release by 2100, 2200, and 2300 from (K. Kretschmer et al., 2015) as a basis, and aggressively assume complete liberation of methane to the atmosphere.

We base radiative forcing changes from reduced Arctic sea ice on work by (Hudson, 2011; Pistone et al., 2019; Nico Wunderling et al., 2020), applying additional radiative inputs reaching 0.3 W/m^2 by 2050 and 0.5 W/m^2 by 2100 for SSP2-4.5 and reaching 0.3 W/m^2 by 2030, 0.6 W/m^2 by 2100, and 0.7 W/m^2 by 2120 for SSP5-8.5, subtracting/adding 15% for low/high scenarios, respectively. This corresponds to an assumed future in which the Arctic sea ice becomes ice-free for a couple summer months after 2100 in SSP2-4.5, and is ice-free throughout the sunlit portion of the year after 2120 in SSP5-8.5. We note that this may somewhat double count existing albedo feedbacks in earth system models, though climatological biases in the Arctic may produce lower sea ice loss and correspondingly weaker albedo feedbacks in many models (Shen, Duan, Li, & Li, 2021).

To account for potential Amazon rainforest ecosystem changes, we started with a current ecosystem carbon stock estimate of 200 Gt C (Cerri et al., 2007; Gibbs et al., 2007; Malhi et al., 2006; S. S. Saatchi et al., 2011), and assume that conversion to savanna or grassland-like vegetation would eventually lead to carbon storage per unit area equivalent to half that of intact tropical rainforest. We further assume that forest areas degraded through edge effects and wildfire exhibit a long-term carbon storage capacity of 25% less than intact forest (Aragão et al., 2018; Silva Junior et al., 2020). Under SSP5-8.5, we assume 60% of Amazonia undergoes conversion from tropical forest to savanna/grassland between 2040 and 2140, with another 20% suffering from degradation due to fire and edge effects. Subsequently, the previously-degraded 20% is converted from tropical forest to savanna/grassland between 2140 and 2200, with another 10% of previously intact forest suffering fire/edge degradation over this period. For SSP-4.5, we assume 30% of the region's tropical forest is converted to savanna/grassland between 2040 and 2100, another 20% suffering degradation due to fire/edge effects. The previously degraded 20% is converted from tropical forest to savanna/grassland between 2100 and 2150, with another 10% of previously intact forest suffering from fire/edge degradation during this time.

We chose to omit potential carbon fluxes or radiative forcing impacts from other candidate tipping elements (boreal forests, stratocumulus cloud decks, tropical monsoons, AMOC, and Greenland/Antarctic ice sheets) on the basis of higher uncertainty surrounding these mechanisms and their potential impacts upon carbon cycling and planetary radiative balance under different warming scenarios. While there is a good probability that one or more of these tipping elements will add net contributions to warming, the current level of scientific knowledge and confidence isn't sufficient to formulate assumptions that aren't largely arbitrary. We do note that around the end of the current century, atmospheric CO_2 concentrations under SSP5-8.5 would reach levels sufficient to risk triggering stratocumulus cloud deck evaporation. However, this proposed climate feedback remains a novel hypothesis with substantial uncertainties, and so we opt to omit this mechanism from the simple model analysis. We further assess the effect of continued die-off of tropical coral reefs to produce no significant global climate feedbacks.

In total, these climate tipping elements likely will not produce large additional warming of several degrees C between the present day and 2300. Rather, the total end-of-century additional warming resulting from the collective contribution of these tipping elements totals ~0.24°C (low: 0.16°C, high: 0.39°C) under SSP2-4.5 and ~0.76°C (low: 0.54°C, high: 1.00°C) under SSP5-8.5 relative to the original scenarios alone (Figure 8). Even employing aggressive estimates for carbon release and radiative forcing changes, the sum effect of these near-term tipping elements is significant but secondary to the larger emissions trajectory. Overall, anthropogenic emissions and the range of potential climate sensitivity are likely to remain the dominant factors relative to these tipping elements in determining total warming over the next few centuries, with other carbon cycle feedbacks serving as a further source of uncertainty. Tipping elements carry substantial implications for carbon budgets and do accelerate rates of warming, but the potential for these mechanisms to drive the transition to a hothouse climate state within centuries remains unclear at present. A strong need nevertheless exists for more rigorous modeling work to further explore possible scenarios, particularly in conjunction with general carbon cycle feedbacks and their associated uncertainties.

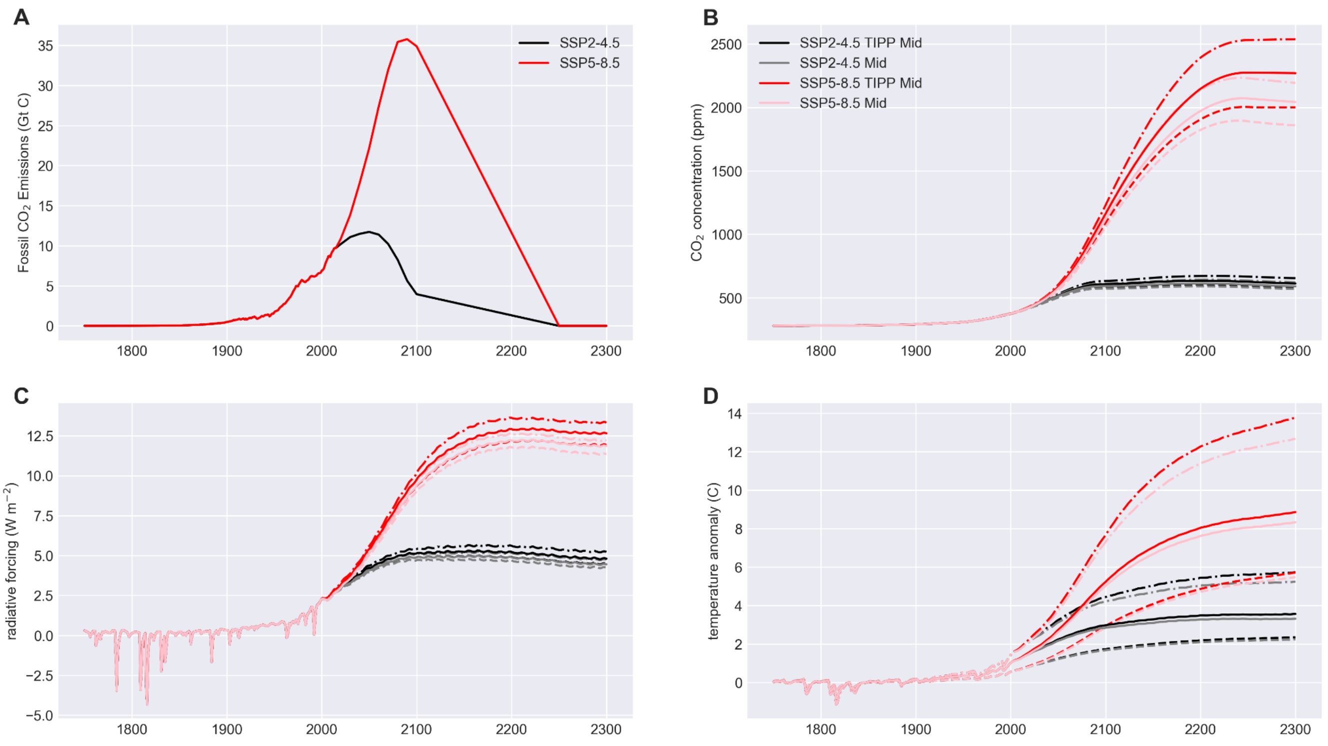

Figure 8: a) fossil fuel CO_2 emissions, b) atmospheric CO_2 concentrations, c) total radiative forcing, and d) global mean surface temperature anomaly relative to pre-industrial (1861–1880 mean) temperatures, plotted from 1750-2300 for a baseline SSP2-4.5 (light gray) and SSP5-8.5 emissions scenario (light red) as well as for modified SSP2-4.5 TIPP (black) and SSP5-8.5 TIPP (red) scenarios incorporating additional CO_2 emissions, CH_4 emissions, and radiative forcing as a result of tipping elements (see Table 5). "Middle" scenarios are plotted using solid lines, while "high" and "low" scenarios are denoted with dot-dashed and dashed lines, respectively.

4.3 Longer-term tipping elements

For tipping elements that act over centuries to millennia, long timescales of action introduce the possibility that in the distant future atmospheric greenhouse gas concentrations may have begun to decline thanks to natural carbon sinks, mitigating overall climate effects. Multi-millennial modeling of several high-forcing scenarios indicates that CO_2 levels ultimately decline by hundreds of parts per million over the course of one to two millennia, driven by natural carbon sinks (Peter U. Clark et al., 2016).

An extended time horizon also leaves a significant opportunity for human society to further drive greenhouse gas removal from the atmosphere through carbon removal methods over future centuries. Many of the future emission scenarios that limit warming to 2°C or below by 2100 already include large-scale use of negative emissions technologies, and the use of these could logically be extended and expanded in following centuries (Riahi et al., 2017). Such considerations may ameliorate the potential impact of tipping elements with higher triggering thresholds that act over extremely long timescales of centuries or more, such as permafrost carbon release (see Section 2.4) and methane hydrate release (Section 2.2). The cumulative effects of permafrost thaw and methane hydrate destabilization manifest over centuries and millennia, respectively (David Archer et al., 2009; McGuire et al., 2018; E. A.G. Schuur et al., 2015; M. Turetsky et al., 2020). More recent research indicates that methane hydrate deposits in particular will likely play a much more minimal role in global climate over coming centuries and millennia than previously believed, as reviewed by (Carolyn D. Ruppel & Kessler, 2017).

Large-scale collapse of the Greenland and West Antarctic ice-sheets (Section 2.3) could strengthen climate change by increasing temperatures through the ice-albedo feedback. However, the planetary albedo change resulting from ice loss takes place on very slow timescales and is additionally balanced by seasonal snow and sea ice and the substantial marine ice generated from ice-sheet collapse itself. Some have even suggested that meltwater inputs from Antarctic ice-sheet loss could even impart a near-term cooling effect on global climate, although understanding of meltwater feedbacks on climate remains incomplete and in need of further study (Bronselaer et al., 2018). Complete loss of land-based ice-sheets in Greenland is anticipated to require multiple millennia (Pattyn et al., 2018), while loss of the Western Antarctic Ice-sheet similarly takes place over a time scale of centuries at minimum even under extreme scenarios (DeConto et al., 2021; Edwards et al., 2021; N. R. Golledge et al., 2015; Nicholas R. Golledge et al., 2019). While some studies have theorized that the GIS and WAIS could serve as initiators for tipping element interactions (N Wunderling et al., 2021), such work has often abstractly represented varying timescales for system response, which in fact play critical roles in determining the likelihood, direction, and significance of interactions.

The likelihood of an AMOC shutdown (Section 2.1) is currently assessed as very low (Collins et al., 2019), with projections still inconclusive as to how much the AMOC may weaken in response to climate change. Current climate models utilize a relatively simple representation of the AMOC and may exhibit excessive AMOC stability (W. Liu et al., 2017), while new oceanographic measurements continue to highlight limitations in our understanding of this system (M. S. Lozier et al., 2019) that impact confidence in model results. Apart from such uncertainties, it also remains unclear whether the larger climate impacts from a weakening AMOC would produce additional net global warming.

(Lenton et al., 2019) additionally invoke the possibility of triggering stratocumulus cloud deck evaporation (Section 2.7) as a potentially important tipping point. However, the level of greenhouse gas forcing required to cause marine stratocumulus cloud formations to break down may lie at around 1200 ppm CO_2 (Schneider et al., 2019), a concentration not attained until towards the end of this century in aggressive, worst-case emissions pathways (Hausfather & Peters, 2020). This threshold remains uncertain, with this mechanism posing sufficient risk to warrant considerable additional research, while emphasizing the value of rapid anthropogenic greenhouse gas emissions reductions.

Overall, many of the long-term tipping elements discussed above exert climate impacts gradually over lengthy timeframes, while other factors such as an AMOC collapse, stratocumulus cloud deck evaporation, or destabilization of methane clathrate deposits may represent high-risk but uncertain or low-probability outcomes. The hypothesis that multiple tipping elements can cascade resulting in several degrees of additional warming remains in need of additional substantiation through quantitative modeling. Nevertheless, the risks of a tipping element cascade and long-term alteration of the Earth's climate trajectory are sufficiently large that any future updates in our understanding of individual tipping elements should motivate reassessment of the dangers of significant additional climate change.

Candidate tipping element	Level of scientific understanding	Predictability by models	Key critical thresholds, global mean warming or otherwise.	Climate impact	Timescale of impacts
AMOC weakening/collapse	Moderate	Good agreement, significant model limitations	Uncertain	Weakening causes regional cooling, wind, precipitation, sea-level changes. Collapse would greatly magnify these impacts.	Weakening occurs over centuries. Collapse would be abrupt or fast.
Methane hydrate destabilization	Moderate	Low	Uncertain, long-term impacts higher beyond ~3°C	Gradual long-term release of carbon to the Earth system	Centuries to multiple millennia
Greenland and Antarctic ice-sheet loss	Moderate	Moderate	Uncertain Greenland: 1.5-5°C West Antarctica: 1.5-3°C East Antarctica: Very uncertain	Multi-meter sea-level rise over centuries to millennia, irreversible ice-sheet loss	Centuries to millennia
Permafrost carbon release	Moderate	Moderate to low	No firm threshold behavior	Added emissions of carbon dioxide and methane	Years to decades, continuing for centuries
Boreal forest ecosystem shifts	Low	Low	Uncertain	Increase in wildfires, significant changes in soil and biomass carbon storage, regional albedo	Decades to a century

				changes, major ecosystem shifts	
Stratocumulus cloud deck evaporation	Low	Low	~1 200 ppm CO_2e	Worldwide marine cloud deck breakup triggers global warming of up to 8°C	Unclear but might be abrupt, with impacts within a decade
Coral reef habitat collapse	Very high	High	Increasingly severe impacts beyond 1.5°C	Degradation of ~99% of warm-water coral habitats worldwide, major socioeconomic impacts	Decades
Amazon rainforest dieback	Good	Moderate	40% deforestation, ~3-4°C, 40% precipitation decrease, or some combination	Die-off of significant fractions of Amazon rainforest, intensification of wildfires, large ecosystem shifts, significant carbon emissions	Decades to a century
Abrupt transitions in S. Asian, African monsoon regime	Low	Moderate to low, but majority of models predict increase in monsoon rainfall.	Thought to be non-tipping element. Increase in regional albedo over India to 0.5 proposed by some.	Gradual increase in mean seasonal precipitation across the monsoon domain.	Decades.
Loss of summer Arctic sea ice	High	Moderate to high	Summer ice loss scales linearly with temperature. Consistently ice-free summers likely for warming of ~2°C.	Increased occurrence of ice-free summer Arctic, global warming feedback	Decades
Tipping element cascade	Low	Low	Uncertain	Significant additional global warming due to interacting climate tipping elements	Centuries to millennia

Table 6: Summary of scientific understanding, key thresholds, and impacts associated with the candidate tipping elements covered in this review.

5 Conclusion

Considering the tipping elements reviewed here as an ensemble (Table 6), some useful commonalities emerge that highlight shared characteristics between some of these systems as well as important priorities for guiding future research. Relatively few of the tipping elements covered in this review are thought to demonstrate a potential for abrupt (within a couple of decades) substantial change that would categorize them as “Gladwellian” tipping elements. Some tipping elements, such as summer Arctic sea ice decline or projected monsoon changes, are arguably more accurately characterized as nontipping elements due to their more predictable response to increasing climate forcing (Boos & Storelvmo, 2016a; Notz & Stroeve, 2016).

Several of the most frequently-invoked candidate climate tipping elements - permafrost thaw (McGuire et al., 2018; M. Turetsky et al., 2020), methane hydrate release (David Archer et al., 2009; K. Kretschmer et al., 2015; Mestdagh et al., 2017), and ice-sheet collapse (DeConto & Pollard, 2016; DeConto et al., 2021; Edwards et al., 2021) - cumulatively act on longer timescales of centuries to millennia, as opposed to driving rapid additional warming or sea-level rise in a matter of years to decades. In the case of other tipping elements (AMOC shutdown/slowdown, methane hydrate release), the likelihood of occurrence and magnitude of impacts under different climate forcing scenarios remains unclear and a topic of continued active research (Good et al., 2018; Carolyn D. Ruppel & Kessler, 2017). For the northern boreal forests, complex, competing feedbacks and ecological interactions complicate efforts to assess overall climate impacts. Nevertheless, a number of tipping elements or climate feedbacks possess both a high certainty of occurrence this century and a relatively strong underlying scientific understanding: Arctic summer sea ice loss, coral reef collapse, and Amazon forest dieback, as well as climatically signficant carbon emissions from thawed permafrost.

The stratocumulus cloud deck evaporation hypothesis and the potential for AMOC slowdown/collapse sit in a unique space. Stratocumulus cloud deck evaporation remains relatively understudied and may require very high concentrations of greenhouse gases to initiate (Schneider et al., 2019). Given the large potential climate impacts of stratocumulus cloud deck breakup, however, constraining the likelihood of and triggering conditions for this powerful climate feedback deserves focused additional research. At the same time, the current state of understanding regarding the response of AMOC to climate change remains incomplete (Collins et al., 2019; Good et al., 2018), and the importance of this potential climate impact to regional and global climate similarly demands further attention.

Overall, tipping elements will play a climatically significant role over the course of the 21st century, but the possibility of a "tipping point cascade" driving a couple or more degrees of additional warming over the next couple centuries remains unlikely. Simultaneously, this review points towards considerable remaining knowledge gaps associated with estimating the probability and impacts of many climate tipping elements. A continued need exists in particular for detailed modeling of the potential influence of tipping elements and carbon cycle feedbacks upon global climate. As scientific understanding of these earth systems and associated feedback mechanisms improves further, the potential for abrupt climate change and for tipping element interactions will require ongoing re-assessment.

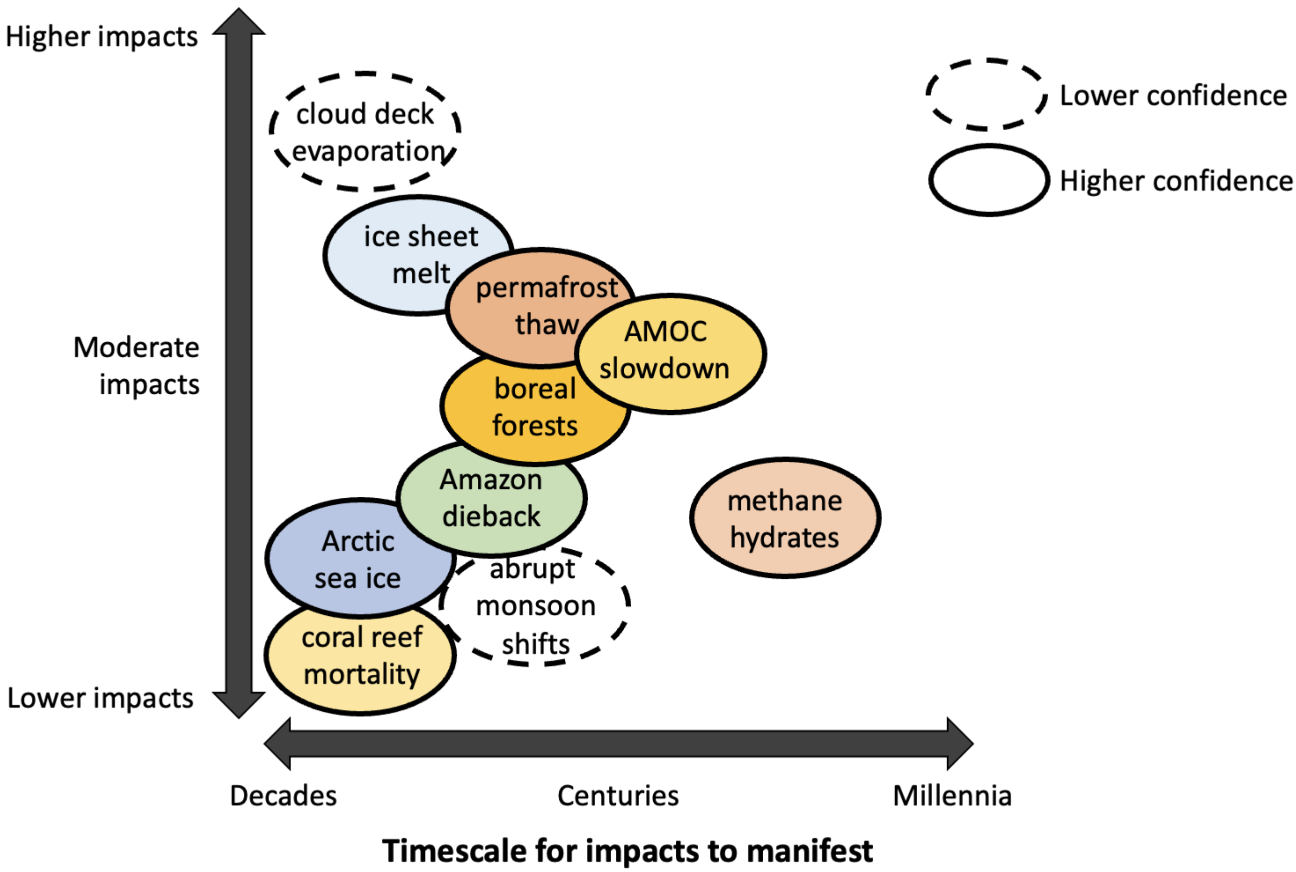

Figure 9: Qualitative two-dimensional organization of individual candidate tipping elements covered in this report. Candidate tipping element mechanisms are organized vertically according to expected impacts on global climate (y-axis) and horizontally based on the timeframe over which those effects will likely manifest. Candidate tipping elements are qualitatively categorized as lower confidence (dashed outline) or higher confidence (solid outline). Colors are arbitrarily assigned for visual presentation. .

This review highlights three tipping elements that pose a more imminent threat of climatically significant carbon release (Figure 9), namely boreal forest ecosystem shifts, permafrost carbon release, and Amazon forest loss. Both the tropical Amazon biome as well as the forests of the boreal north represent large repositories of carbon potentially vulnerable over the next century to large-scale ecosystem shifts (Cox et al., 2013; Mcguire et al., 2009). Meanwhile, although the bulk of permafrost carbon release will take place over centuries, climatically significant quantities of carbon could be liberated from permafrost in the near term (M. Turetsky et al., 2020). Both the Amazon and the Arctic are ecosystems currently under significant stress that could trigger committed release of climate-altering quantities of greenhouse gases, yet substantial uncertainties remain. These yet-unresolved questions fundamentally control how much climate change the Amazon rainforest and northern Arctic could independently drive. With the fate of biomass carbon in both ecosystems also closely tied to human land management practices, forest conservation policy in Brazil, Russia, Canada, Scandinavia, China and north central Asia may carry weighty climate implications. Both the Amazon as well as the circumpolar north consequently represent key priorities for ongoing research.

The current literature also suggests that most of the earth systems covered here are not yet committed to fixed, substantial changes in response to warming. “Slow-onset” tipping points may exhibit some resiliency due to the longer timescales over which they respond to climate forcing, potentially allowing for recovery to the original system state if overshoots of critical thresholds are smaller and/or limited in duration (Ritchie et al., 2021). These considerations emphasize the significant potential for climate mitigation to moderate the impacts of tipping elements. Modeling of carbon release and other impacts associated with tipping elements such as ice-sheet loss, Arctic sea ice decline, permafrost carbon release, and AMOC weakening point towards large differences between worst-case and emissions pathways featuring stronger climate mitigation efforts. Limiting deforestation is also understood to play a key role in maximizing the long-term health of the Amazon tropical forest (Carlos A. Nobre et al., 2016). Decisive actions to cut greenhouse gas emissions and reduce land-use impacts, potentially in combination with deployment of negative emissions technologies at scale, can thus dramatically reduce the future consequences associated with many tipping elements.

Nevertheless, tipping elements presently remain one of the larger unknowns involved in predicting future warming and climate impacts. Continuing to better constrain the range of uncertainty surrounding likelihood and impacts of climate tipping elements represents a valuable ongoing contribution of the earth science community, benefiting policymakers and planners seeking to assess risks and craft effective climate policies moving forwards. For the foreseeable future, research into climate tipping elements will consequently remain of high value and in high demand both within and beyond the research community.

Code availability

Python code and data files needed to replicate our simple FaIR modeling exercise are available at Zenodo using the following DOI: 10.5281/zenodo.5120052.

Author contributions

SW was responsible for initial literature review, core writing, and figure generation for the article. Contributions to the writing and editing of individual topic sections were as follows. AMOC: SZ, WL, and SW. Methane hydrates: JK and SW. Ice sheets: SW. Permafrost: MT and SW. Boreal forests: AF, MT, and SW. Monsoons: WB and SW. Stratocumulus cloud decks: SW. Coral reefs: EL and SW. Amazon rainforest: LA and SW. Arctic sea ice: JS and SW. Both SW and ZH worked to produce the illustrative simple climate model exercise with input from AF, MT, LA, RK, and WB. All authors contributed to review and editing of the synthesis sections, figures, and tables included in the manuscript.

Acknowledgements

No dedicated grants funded this research. Funding sources for the Breakthrough Institute can be found here: https://thebreakthrough.org/about/who-we-are/funders. We thank Nicholas Foukal, David Archer, Olivier Gagliardini, Ted Schuur, Dirk Notz, Laifang Li, Deepti Singh, Daniel Nepstad, Laura Borma, Heidi Hirsch, Tapio Schneider, and Jacquelyn Shuman for their valuable feedback on sections of an earlier draft. We also thank Tamsin Edwards and Robert Kopp for valuable input on the manuscript as well as Paul Baker, Patrick Brown, and Brad Murray for helpful conversations. We are grateful to Nicholas Leach, Chris Smith, and Zebedee Nicholls for assistance and input in the FaIR v.1.6.3 modeling exercise.

Competing interests

The authors declare no conflict of interest.

References

Abbott, B. W., Jones, J. B., Godsey, S. E., Larouche, J. R., & Bowden, W. B. (2015). Patterns and persistence of hydrologic carbon and nutrient export from collapsing upland permafrost. *Biogeosciences*, *12*(12), 3725–3740. https://doi.org/10.5194/bg-12-3725-2015

Abbott, Benjamin W., & Jones, J. B. (2015). Permafrost collapse alters soil carbon stocks, respiration, CH4, and N2O in upland tundra. *Global Change Biology*, *21*(12), 4570–4587. https://doi.org/10.1111/gcb.13069

Adkins, J., deMenocal, P., & Eshel, G. (2006). The "African humid period" and the record of marine upwelling from excess 230Th in Ocean Drilling Program Hole 658C. *Paleoceanography*, *21*(4). https://doi.org/https://doi.org/10.1029/2005PA001200

Albano, L. J., Turetsky, M. R., Mack, M. C., & Kane, E. S. (2021). Deep roots of Carex aquatilis have greater ammonium uptake capacity than shallow roots in peatlands following permafrost thaw. *Plant and Soil*. https://doi.org/10.1007/s11104-021-04978-x

Alexander, H. D., & Mack, M. C. (2016). A Canopy Shift in Interior Alaskan Boreal Forests: Consequences for Above- and Belowground Carbon and Nitrogen Pools during Post-fire Succession. *Ecosystems*, *19*(1), 98–114. https://doi.org/10.1007/s10021-015-9920-7

Alkhayuon, H., Ashwin, P., Jackson, L. C., Quinn, C., & Wood, R. A. (2019). Basin bifurcations, oscillatory instability and rate-induced thresholds for Atlantic meridional overturning circulation in a global oceanic box model. *Proceedings of* *the Royal Society A: Mathematical, Physical and Engineering Sciences*, *475*(2225), 20190051. https://doi.org/10.1098/rspa.2019.0051

Allen, C. D., Macalady, A. K., Chenchouni, H., Bachelet, D., McDowell, N., Vennetier, M., … Cobb, N. (2010). A global overview of drought and heat-induced tree mortality reveals emerging climate change risks for forests. *Forest Ecology and Management*, *259*(4), 660–684. https://doi.org/10.1016/j.foreco.2009.09.001

Alley, R. B., Anandakrishnan, S., & Jung, P. (2001). Stochastic Resonance in the North Atlantic. *Paleoceanography*, *16*(2), 190–198.

Anderson, L. O., Ribeiro Neto, G., Cunha, A. P., Fonseca, M. G., Mendes de Moura, Y., Dalagnol, R., … de Aragão, L. E. O. e C. (2018). Vulnerability of Amazonian forests to repeated droughts. *Philosophical Transactions of the Royal Society B: Biological Sciences*, *373*(1760), 20170411. https://doi.org/10.1098/rstb.2017.0411

Andresen, C. G., Lawrence, D. M., Wilson, C. J., McGuire, A. D., Koven, C., Schaefer, K., … Zhang, W. (2020). Soil moisture and hydrology projections of the permafrost region -- a model intercomparison. *The Cryosphere*, *14*(2), 445–459. https://doi.org/10.5194/tc-14-445-2020

Anton, A., Randle, J. L., Garcia, F. C., Rossbach, S., Ellis, J. I., Weinzierl, M., & Duarte, C. M. (2020). Differential thermal tolerance between algae and corals may trigger the proliferation of algae in coral reefs. *Global Change Biology*, *26*(8), 4316–4327. https://doi.org/https://doi.org/10.1111/gcb.15141

Aragão, L. E. O. C., Anderson, L. O., Fonseca, M. G., Rosan, T. M., Vedovato, L. B., Wagner, F. H., … Saatchi, S. (2018). 21st Century drought-related fires counteract the decline of Amazon deforestation carbon emissions. *Nature Communications*, *9*(1), 536. https://doi.org/10.1038/s41467-017-02771-y

Archer, D. (2007). Methane hydrate stability and anthropogenic climate change. In *European Geosciences Union* (Vol. 4). Retrieved from https://hal.archives-ouvertes.fr/hal-00297882

Archer, D. (2015). A model of the methane cycle, permafrost, and hydrology of the Siberian continental margin. *Biogeosciences*, *12*(10), 2953–2974. https://doi.org/10.5194/bg-12-2953-2015

Archer, David, Buffett, B., & Brovkin, V. (2009). Ocean methane hydrates as a slow tipping point in the global carbon cycle. *Proceedings of the National Academy of Sciences of the United States of America*, *106*(49), 20596–20601. https://doi.org/10.1073/pnas.0800885105

Armour, K. C., Eisenman, I., Blanchard-Wrigglesworth, E., McCusker, K. E., & Bitz, C. M. (2011). The reversibility of sea ice loss in a state-of-the-art climate model. *Geophysical Research Letters*, *38*(16), 1–5. https://doi.org/10.1029/2011GL048739

Årthun, M., Onarheim, I. H., Dörr, J., & Eldevik, T. (2021). The Seasonal and Regional Transition to an Ice-Free Arctic. *Geophysical Research Letters*, *48*(1), e2020GL090825. https://doi.org/https://doi.org/10.1029/2020GL090825

Arvor, D., Dubreuil, V., Ronchail, J., Simões, M., & Funatsu, B. M. (2014). Spatial patterns of rainfall regimes related to levels of double cropping agriculture systems in Mato Grosso (Brazil). *International Journal of Climatology*, *34*(8), 2622–2633. https://doi.org/10.1002/joc.3863

Aubry-Kientz, M., Rossi, V., Cornu, G., Wagner, F., & Hérault, B. (2019). Temperature rising would slow down tropical forest dynamic in the Guiana Shield. *Scientific Reports*, *9*(1), 10235. https://doi.org/10.1038/s41598-019-46597-8

Baker, A. C., Glynn, P. W., & Riegl, B. (2008). Climate change and coral reef bleaching: An ecological assessment of long-term impacts, recovery trends and future outlook. *Estuarine, Coastal and Shelf Science*, *80*(4), 435–471. https://doi.org/10.1016/j.ecss.2008.09.003

Bakker, A. M. R., Louchard, D., & Keller, K. (2017). Sources and implications of deep uncertainties surrounding sea-level projections. *Climatic Change*, *140*(3–4), 339–347. https://doi.org/10.1007/s10584-016-1864-1

Bakker, P., Schmittner, A., Lenaerts, J. T. M., Abe-Ouchi, A., Bi, D., van den Broeke, M. R., … Yin, J. (2016). Fate of the Atlantic Meridional Overturning Circulation: Strong decline under continued warming and Greenland melting. *Geophysical Research Letters*, *43*(23), 12,252-12,260. https://doi.org/10.1002/2016GL070457

Balch, J. K., Brando, P. M., Nepstad, D. C., Coe, M. T., Silvério, D., Massad, T. J., … Carvalho, K. S. (2015). The Susceptibility of Southeastern Amazon Forests to Fire: Insights from a Large-Scale Burn Experiment. *BioScience*, *65*(9), 893–905. https://doi.org/10.1093/biosci/biv106

Balshi, M. S., McGuire, A. D., Duffy, P., Flannigan, M., Walsh, J., & Melillo, J. (2009). Assessing the response of area burned to changing climate in western boreal North America using a Multivariate Adaptive Regression Splines (MARS) approach. *Global Change Biology*, *15*(3), 578–600. https://doi.org/10.1111/j.1365-2486.2008.01679.x

Baltzer, J. L., Veness, T., Chasmer, L. E., Sniderhan, A. E., & Quinton, W. L. (2014). Forests on thawing permafrost: fragmentation, edge effects, and net forest loss. *Global Change Biology*, *20*(3), 824–834. https://doi.org/https://doi.org/10.1111/gcb.12349

Bamber, J. L., Oppenheimer, M., Kopp, R. E., Aspinall, W. P., & Cooke, R. M. (2019). Ice sheet contributions to future sea-level rise from structured expert judgment. *Proceedings of the National Academy of Sciences of the United States of America*, *166*(23), 11195–11200. https://doi.org/10.1073/pnas.1817205116

Bamber, J. L., Westaway, R. M., Marzeion, B., & Wouters, B. (2018). Corrigendum: The land ice contribution to sea level during the satellite era (Environmental Research Letters (2018) 13 (063008) DOI: 10.1088/1748-9326/aac2f0). *Environmental Research Letters*, *13*(9). https://doi.org/10.1088/1748-9326/aadb2c

Barletta, V. R., Bevis, M., Smith, B. E., Wilson, T., Brown, A., Bordoni, A., … Wiens, D. A. (2018). Observed rapid bedrock uplift in amundsen sea embayment promotes ice-sheet stability. *Science*, *360*(6395), 1335–1339. https://doi.org/10.1126/science.aao1447

Barnes, E. A., & Screen, J. A. (2015). The impact of Arctic warming on the midlatitude jet-stream: Can it? Has it? Will it? *WIREs Climate Change*, *6*(3), 277–286. https://doi.org/https://doi.org/10.1002/wcc.337

Bassis, J. N., & Jacobs, S. (2013). Diverse calving patterns linked to glacier geometry. *Nature Geoscience*, *6*(10), 833–836. https://doi.org/10.1038/ngeo1887

Bassis, J. N., & Walker, C. C. (2012). Upper and lower limits on the stability of calving glaciers from the yield strength envelope of ice. *Proceedings of the Royal Society A: Mathematical, Physical and Engineering Sciences*, *468*(2140), 913–931. https://doi.org/10.1098/rspa.2011.0422

Bathiany, S., Notz, D., Mauritsen, T., Raedel, G., & Brovkin, V. (2016). On the potential for abrupt Arctic winter sea ice loss. *Journal of Climate*, *29*(7), 2703–2719. https://doi.org/10.1175/JCLI-D-15-0466.1

Bathiany, S., Scheffer, M., Van Nes, E. H., Williamson, M. S., & Lenton, T. M. (2018). Abrupt Climate Change in an Oscillating World. *Scientific Reports*, *8*(1). https://doi.org/10.1038/s41598-018-23377-4

Beck, P. S. A., Juday, G. P., Alix, C., Barber, V. A., Winslow, S. E., Sousa, E. E., … Goetz, S. J. (2011). Changes in forest productivity across Alaska consistent with biome shift. *Ecology Letters*, *14*(4), 373–379. https://doi.org/10.1111/j.1461-0248.2011.01598.x

Bell, R. E. (2008, May). The role of subglacial water in ice-sheet mass balance. *Nature Geoscience*, Vol. 1, pp. 297–304. https://doi.org/10.1038/ngeo186

Bender, F. A. M., Ramanathan, V., & Tselioudis, G. (2012). Changes in extratropical storm track cloudiness 1983-2008: Observational support for a poleward shift. *Climate Dynamics*, *38*(9–10), 2037–2053. https://doi.org/10.1007/s00382-011-1065-6

Berchet, A., Bousquet, P., Pison, I., Locatelli, R., Chevallier, F., Paris, J.-D., … Hermansen, O. (2016). Atmospheric constraints on the methane emissions from the \hack{\newline}East Siberian Shelf. *Atmospheric Chemistry and Physics*, *16*(6), 4147–4157. https://doi.org/10.5194/acp-16-4147-2016

Berenguer, E., Carvalho, N., Anderson, L. O., Aragão, L. E. O. C., França, F., & Barlow, J. (2021). Improving the spatial-temporal analysis of Amazonian fires. *Global Change Biology*, *27*(3), 469–471. https://doi.org/https://doi.org/10.1111/gcb.15425

Berenguer, E., Ferreira, J., Gardner, T. A., Aragão, L. E. O. C., De Camargo, P. B., Cerri, C. E., … Barlow, J. (2014). A large-scale field assessment of carbon stocks in human-modified tropical forests. *Global Change Biology*, *20*(12), 3713–3726. https://doi.org/https://doi.org/10.1111/gcb.12627

Berger, A., & Loutre, M. F. (2002). An exceptionally long interglacial ahead? *Science*, *297*(5585), 1287–1288. https://doi.org/10.1126/science.1076120

Bertani, G., Wagner, F. H., Anderson, L. O., & Aragão, L. E. O. C. (2017). Chlorophyll Fluorescence Data Reveals Climate-Related Photosynthesis Seasonality in Amazonian Forests. *Remote Sensing*, *9*(12). https://doi.org/10.3390/rs9121275

Betts, R. A. (2000). Offset of the potential carbon sink from boreal forestation by decreases in surface albedo. *Nature*, *408*(6809), 187–190. https://doi.org/10.1038/35041545

Betts, R., Sanderson, M., & Woodward, S. (2008). Effects of large-scale Amazon forest degradation on climate and air quality through fluxes of carbon dioxide, water, energy, mineral dust and isoprene. *Philosophical Transactions of the Royal Society B: Biological Sciences*, *363*(1498), 1873–1880. https://doi.org/10.1098/rstb.2007.0027

Bevis, M., Harig, C., Khan, S. A., Brown, A., Simons, F. J., Willis, M., … Nylen, T. (2019). Accelerating changes in ice mass within Greenland, and the ice sheet's sensitivity to atmospheric forcing. *Proceedings of the National Academy of Sciences of the United States of America*, *116*(6), 1934–1939. https://doi.org/10.1073/pnas.1806562116

Biastoch, A, Treude, T., Rüpke, L. H., Riebesell, U., Roth, C., Burwicz, E. B., … Wallmann, K. (2011). Rising Arctic Ocean temperatures cause gas hydrate destabilization and ocean acidification. *Geophysical Research Letters*, *38*(8). https://doi.org/https://doi.org/10.1029/2011GL047222

Biastoch, Arne, Böning, C. W., Getzlaff, J., Molines, J.-M., & Madec, G. (2008). Causes of Interannual–Decadal Variability in the Meridional Overturning Circulation of the Midlatitude North Atlantic Ocean. *Journal of Climate*, *21*(24), 6599–6615. https://doi.org/10.1175/2008JCLI2404.1

Biasutti, M. (2016). What brings rain to the Sahel? *Nature Climate Change*, *6*(10), 897–898. https://doi.org/10.1038/nclimate3080

Biasutti, M. (2019). Rainfall trends in the African Sahel: Characteristics, processes, and causes. *WIREs Climate Change*, *10*(4), e591. https://doi.org/https://doi.org/10.1002/wcc.591

Biasutti, M., & Giannini, A. (2006). Robust Sahel drying in response to late 20th century forcings. *Geophysical Research Letters*, *33*(11). https://doi.org/https://doi.org/10.1029/2006GL026067

Biasutti, M., Voigt, A., Boos, W. R., Braconnot, P., Hargreaves, J. C., Harrison, S. P., … Xie, S. P. (2018, June 1). Global energetics and local physics as drivers of past, present and future monsoons. *Nature Geoscience*, Vol. 11, pp. 392–400. https://doi.org/10.1038/s41561-018-0137-1

Bindoff, N. L., Cheung, W. W. L., Kairo, J. G., Arístegui, J., Guinder, V. A., Hallberg, R., … Williamson, P. (2019). Changing Oean, Marine Ecosystems, and Dependent Communities. In *IPCC Special Report on the Ocean and Cryosphere in a Changing Climate*. https://doi.org/https://www.ipcc.ch/report/srocc/

Bingham, R. J., Hughes, C. W., Roussenov, V., & Williams, R. G. (2007). Meridional coherence of the North Atlantic meridional overturning circulation. *Geophysical Research Letters*, *34*(23). https://doi.org/https://doi.org/10.1029/2007GL031731

Biskaborn, B. K., Smith, S. L., Noetzli, J., Matthes, H., Vieira, G., Streletskiy, D. A., … Lantuit, H. (2019). Permafrost is warming at a global scale. *Nature Communications*, *10*(1). https://doi.org/10.1038/s41467-018-08240-4

Bitz, C. M., Fyfe, J. C., & Flato, G. M. (2002). Sea ice response to wind forcing from AMIP models. *Journal of Climate*, *15*(5), 522–536. https://doi.org/10.1175/1520-0442(2002)015<0522:SIRTWF>2.0.CO;2

Bitz, C. M., & Roe, G. H. (2004). A mechanism for the high rate of sea ice thinning in the Arctic Ocean. *Journal of Climate*, *17*(18), 3623–3632. https://doi.org/10.1175/1520-0442(2004)017<3623:AMFTHR>2.0.CO;2

Bjorkman, A. D., Myers-Smith, I. H., Elmendorf, S. C., Normand, S., Rüger, N., Beck, P. S. A., … Weiher, E. (2018). Plant functional trait change across a warming tundra biome. *Nature*, *562*(7725), 57–62. https://doi.org/10.1038/s41586-018-0563-7

Blossey, P. N., Bretherton, C. S., Zhang, M., Cheng, A., Endo, S., Heus, T., … Xu, K. M. (2013). Marine low cloud sensitivity to an idealized climate change: The CGILS les intercomparison. *Journal of Advances in Modeling Earth Systems*, *5*(2), 234–258. https://doi.org/10.1002/jame.20025

Boening, C., Lebsock, M., Landerer, F., & Stephens, G. (2012). Snowfall-driven mass change on the East Antarctic ice sheet.

Geophysical Research Letters, *39*(21), n/a-n/a. https://doi.org/10.1029/2012GL053316

Bollasina, M. A., Ming, Y., & Ramaswamy, V. (2011). Anthropogenic Aerosols and the Weakening of the South Asian Summer Monsoon. *Science*, *334*(6055), 502 LP – 505. https://doi.org/10.1126/science.1204994

Bonan, G. B., Pollard, D., & Thompson, S. L. (1992). Effects of boreal forest vegetation on global climate. *Nature*, *359*(6397), 716–718. https://doi.org/10.1038/359716a0

Bonnaventure, P. P., Smith, S. L., Lamoureux, S. F., Way, R. G., Ednie, M., Bouchard, F., … Godin, E. (2018). Permafrost. In *From Science to Policy in the Eastern Canadian Arctic: An Integrated Regional Impact Study (IRIS) of Climate Change and Modernization* (pp. 119–139). Québec City, Canada: ArcticNet.

Boos, W. R., & Emanuel, K. A. (2009). Annual intensification of the Somali jet in a quasi-equilibrium framework: Observational composites. *Quarterly Journal of the Royal Meteorological Society*, *135*(639), 319–335. https://doi.org/https://doi.org/10.1002/qj.388

Boos, W. R., & Korty, R. L. (2016). Regional energy budget control of the intertropical convergence zone and application to mid-Holocene rainfall. *Nature Geoscience*, *9*(12), 892–897. https://doi.org/10.1038/ngeo2833

Boos, W. R., & Storelvmo, T. (2016a). Near-linear response of mean monsoon strength to a broad range of radiative forcings. *Proceedings of the National Academy of Sciences of the United States of America*, *113*(6), 1510–1515. https://doi.org/10.1073/pnas.1517143113

Boos, W. R., & Storelvmo, T. (2016b, April 26). Linear scaling for monsoons based on well-verified balance between adiabatic cooling and latent heat release. *Proceedings of the National Academy of Sciences of the United States of America*, Vol. 113, pp. E2350–E2351. https://doi.org/10.1073/pnas.1603626113

Bopp, L., Resplandy, L., Orr, J. C., Doney, S. C., Dunne, J. P., Gehlen, M., … Vichi, M. (2013). Multiple stressors of ocean ecosystems in the 21st century: Projections with CMIP5 models. *Biogeosciences*, *10*(10), 6225–6245. https://doi.org/10.5194/bg-10-6225-2013

Boswell, R., & Collett, T. S. (2011, April). Current perspectives on gas hydrate resources. *Energy and Environmental Science*, Vol. 4, pp. 1206–1215. https://doi.org/10.1039/c0ee00203h

Boudreau, B. P., Luo, Y., Meysman, F. J. R., Middelburg, J. J., & Dickens, G. R. (2015). Gas hydrate dissociation prolongs acidification of the Anthropocene oceans. *Geophysical Research Letters*, *42*(21), 9337-9344A. https://doi.org/https://doi.org/10.1002/2015GL065779

Bove, C. B., Ries, J. B., Davies, S. W., Westfield, I. T., Umbanhowar, J., & Castillo, K. D. (2019). Common Caribbean corals exhibit highly variable responses to future acidification and warming. *Proceedings of the Royal Society B: Biological Sciences*, *286*(1900). https://doi.org/10.1098/rspb.2018.2840

Bowen, J. C., Ward, C. P., Kling, G. W., & Cory, R. M. (2020). Arctic Amplification of Global Warming Strengthened by Sunlight Oxidation of Permafrost Carbon to CO2. *Geophysical Research Letters*, *47*(12), e2020GL087085. https://doi.org/https://doi.org/10.1029/2020GL087085

Boyd, M. A., Berner, L. T., Foster, A. C., Goetz, S. J., Rogers, B. M., Walker, X. J., & Mack, M. C. (2021). Historic declines in growth portend trembling aspen death during a contemporary leaf miner outbreak in Alaska. *Ecosphere*, *12*(6), e03569. https://doi.org/https://doi.org/10.1002/ecs2.3569

Bradshaw, C. J. A., & Warkentin, I. G. (2015). Global estimates of boreal forest carbon stocks and flux. *Global and Planetary Change*, *128*, 24–30. https://doi.org/10.1016/j.gloplacha.2015.02.004

Brando, P.M, Soares-Filho, B., Rodrigues, L., Assunção, A., Morton, D., Tuchschneider, D., … Coe, M. . (2019). The gathering firestorm in southern Amazonia. *Science Advance in Press.*, (January), 1–10.

Brando, Paulo Monteiro, Balch, J. K., Nepstad, D. C., Morton, D. C., Putz, F. E., Coe, M. T., … Soares-Filho, B. S. (2014). Abrupt increases in Amazonian tree mortality due to drought-fire interactions. *Proceedings of the National Academy of Sciences of the United States of America*, *111*(17), 6347–6352. https://doi.org/10.1073/pnas.1305499111

Bretherton, C. S. (2015). Insights into low-latitude cloud feedbacks from high-resolution models. *Philosophical Transactions of the Royal Society A: Mathematical, Physical and Engineering Sciences*, *373*(2054). https://doi.org/10.1098/rsta.2014.0415

Bretherton, C. S., & Wyant, M. C. (1997). Moisture transport, lower-tropospheric stability, and decoupling of cloud-topped boundary layers. *Journal of the Atmospheric Sciences*, *54*(1), 148–167. https://doi.org/10.1175/1520-0469(1997)054<0148:MTLTSA>2.0.CO;2

Brodribb, T. J., Cochard, H., & Dominguez, C. R. (2019). Themed Issue Article: Climate Change Impact on Urban and Natural Measuring the pulse of trees ; using the vascular system to predict tree mortality in the 21st century. *Conservation Physiology*, *7*, 1–7. https://doi.org/10.1093/conphys/coz046..Rising

Brodribb, T. J., Powers, J., Cochard, H., & Choat, B. (2020). Hanging by a thread? Forests and drought. *Science*, *368*(6488), 261–266. https://doi.org/10.1126/science.aat7631

Broecker, W.S. (1991). The great ocean conveyor. *Oceanography*. https://doi.org/10.1016/s0262-4079(08)61198-7

Broecker, Wallace S, Peteet, D. M., & Rind, D. (1985). Does the ocean–atmosphere system have more than one stable mode of operation? *Nature*, *315*(6014), 21–26. https://doi.org/10.1038/315021a0

Bronselaer, B., Winton, M., Griffies, S. M., Hurlin, W. J., Rodgers, K. B., Sergienko, O. V., … Russell, J. L. (2018). Change in future climate due to Antarctic meltwater. *Nature*, *564*(7734), 53–58. https://doi.org/10.1038/s41586-018-0712-z

Brown, B. E., Dunne, R. P., Goodson, M. S., & Douglas, A. E. (2002). Experience shapes the susceptibility of a reef coral to bleaching. *Coral Reefs*, *21*(2), 119–126. https://doi.org/10.1007/s00338-002-0215-z

Bruno, J. F., & Selig, E. R. (2007). Regional decline of coral cover in the Indo-Pacific: Timing, extent, and subregional comparisons. *PLoS ONE*, *2*(8). https://doi.org/10.1371/journal.pone.0000711

Bryden, H. L., Johns, W. E., King, B. A., McCarthy, G., McDonagh, E. L., Moat, B. I., & Smeed, D. A. (2020). Reduction in ocean heat transport at 26°N since 2008 cools the eastern subpolar gyre of the North Atlantic Ocean. *Journal of Climate*, *33*(5), 1677–1689. https://doi.org/10.1175/JCLI-D-19-0323.1

Bryden, H. L., Longworth, H. R., & Cunningham, S. A. (2005). Slowing of the Atlantic meridional overturning circulation at 25° N. *Nature*, *438*(7068), 655–657. https://doi.org/10.1038/nature04385

Buckley, M. W., Ferreira, D., Campin, J. M., Marshall, J., & Tulloch, R. (2012). On the relationship between decadal buoyancy anomalies and variability of the atlantic meridional overturning circulation. *Journal of Climate*, *25*(23), 8009–8030. https://doi.org/10.1175/JCLI-D-11-00505.1

Buckley, M. W., & Marshall, J. (2016, March 1). Observations, inferences, and mechanisms of the Atlantic Meridional Overturning Circulation: A review. *Reviews of Geophysics*, Vol. 54, pp. 5–63. https://doi.org/10.1002/2015RG000493

Bullock, E. L., & Woodcock, C. E. (2020). Carbon loss and removal due to forest disturbance and regeneration in the Amazon. *Science of The Total Environment*, 142839. https://doi.org/https://doi.org/10.1016/j.scitotenv.2020.142839

Bullock, E. L., Woodcock, C. E., Souza Jr., C., & Olofsson, P. (2020). Satellite-based estimates reveal widespread forest degradation in the Amazon. *Global Change Biology*, *26*(5), 2956–2969. https://doi.org/https://doi.org/10.1111/gcb.15029

Bulthuis, K., Arnst, M., Sun, S., & Pattyn, F. (2019). Uncertainty quantification of the multi-centennial response of the antarctic ice sheet to climate change. *Cryosphere*, *13*(4), 1349–1380. https://doi.org/10.5194/tc-13-1349-2019

Burke, E. J., Zhang, Y., & Krinner, G. (2020). Evaluating permafrost physics in the Coupled Model Intercomparison Project 6 (CMIP6) models and their sensitivity to climate change. *The Cryosphere*, *14*(9), 3155–3174. https://doi.org/10.5194/tc-14-3155-2020

Burke, K. D., Williams, J. W., Chandler, M. A., Haywood, A. M., Lunt, D. J., & Otto-Bliesner, B. L. (2018). *Pliocene and Eocene provide best analogs for near-future climates*. https://doi.org/10.5061/dryad.0j18k00

Burke, L., Reytar, K., Spalding, M., & Perry, A. (2011). Reefs at Risk Revisited. In *Defenders* (Vol. 74). Retrieved from http://www.pubmedcentral.nih.gov/articlerender.fcgi?artid=3150666&tool=pmcentrez&rendertype=abstract

Burrell, A., Kukavskaya, E., Baxter, R., Sun, Q., & Barrett, K. (2021). *Post-fire Recruitment Failure as a Driver of Forest to Non-forest Ecosystem Shifts in Boreal Regions BT - Ecosystem Collapse and Climate Change* (J. G. Canadell & R. B. Jackson, Eds.). https://doi.org/10.1007/978-3-030-71330-0_4

Caballero, R., & Huber, M. (2013). State-dependent climate sensitivity in past warm climates and its implications for future climate projections. *Proceedings of the National Academy of Sciences*, *110*(35), 14162 LP – 14167. https://doi.org/10.1073/pnas.1303365110

Caesar, L., McCarthy, G. D., Thornalley, D. J. R., Cahill, N., & Rahmstorf, S. (2021). Current Atlantic Meridional Overturning Circulation weakest in last millennium. *Nature Geoscience*, *14*(3), 118–120. https://doi.org/10.1038/s41561-021-00699-z

Caesar, L., Rahmstorf, S., Robinson, A., Feulner, G., & Saba, V. (2018). Observed fingerprint of a weakening Atlantic Ocean overturning circulation. *Nature*, *556*(7700), 191–196. https://doi.org/10.1038/s41586-018-0006-5

Cai, Y., Lenton, T. M., & Lontzek, T. S. (2016). Risk of multiple interacting tipping points should encourage rapid CO2 emission reduction. *Nature Climate Change*, *6*(5), 520–525. https://doi.org/10.1038/nclimate2964

Carpino, O. A., Berg, A. A., Quinton, W. L., & Adams, J. R. (2018). Climate change and permafrost thaw-induced boreal forest loss in northwestern Canada. *Environmental Research Letters*, *13*(8). https://doi.org/10.1088/1748-9326/aad74e

Carroll, D., Sutherland, D. A., Hudson, B., Moon, T., Catania, G. A., Shroyer, E. L., … van den Broeke, M. R. (2016). The impact of glacier geometry on meltwater plume structure and submarine melt in Greenland fjords. *Geophysical Research Letters*, *43*(18), 9739–9748. https://doi.org/10.1002/2016GL070170

Cerri, C. E. P., Easter, M., Paustian, K., Killian, K., Coleman, K., Bernoux, M., … Cerri, C. C. (2007). Predicted soil organic carbon stocks and changes in the Brazilian Amazon between 2000 and 2030. *Agriculture, Ecosystems and Environment*, *122*(1), 58–72. https://doi.org/10.1016/j.agee.2007.01.008

Chadburn, S. E., Burke, E. J., Cox, P. M., Friedlingstein, P., Hugelius, G., & Westermann, S. (2017). An observation-based constraint on permafrost loss as a function of global warming. *Nature Climate Change*, *7*(5), 340–344. https://doi.org/10.1038/nclimate3262

Chan, E. W., Shiller, A. M., Joung, D. J., Arrington, E. C., Valentine, D. L., Redmond, M. C., … Kessler, J. D. (2019). Investigations of Aerobic Methane Oxidation in Two Marine Seep Environments: Part 1—Chemical Kinetics. *Journal of Geophysical Research: Oceans*, *124*(12), 8852–8868. https://doi.org/https://doi.org/10.1029/2019JC015594

Chapin, F. S., Sturm, M., Serreze, M. C., McFadden, J. P., Key, J. R., Lloyd, A. H., … Welker, J. M. (2005). Role of land-surface changes in arctic summer warming. *Science*, *310*(5748), 657–660. https://doi.org/10.1126/science.1117368

Charney, J G. (1975). Dynamics of deserts and drought in the Sahel. *Quarterly Journal of the Royal Meteorological Society*, *101*(428), 193–202. https://doi.org/https://doi.org/10.1002/qj.49710142802

Charney, Jule G, & Eliassen, A. (1964). On the Growth of the Hurricane Depression. *Journal of Atmospheric Sciences*, *21*(1), 68–75. https://doi.org/10.1175/1520-0469(1964)021<0068:OTGOTH>2.0.CO;2

Chen, Z., Zhou, T., Zhang, L., Chen, X., Zhang, W., & Jiang, J. (2020). Global Land Monsoon Precipitation Changes in CMIP6 Projections. *Geophysical Research Letters*, *47*(14), e2019GL086902. https://doi.org/https://doi.org/10.1029/2019GL086902

Christensen, J. H., Kanikicharla, K. K., Aldrian, E., An, S. Il, Cavalcanti, I. F. A., de Castro, M., … Zhou, T. (2013). Climate phenomena and their relevance for future regional climate change. In *Climate Change 2013: the Physical Science Basis. Working Group I Contribution to the Fifth Assessment Report of the Intergovernmental Panel on Climate Change*. https://doi.org/10.1017/CBO9781107415324.028

Circone, S., Kirby, S. H., & Stern, L. A. (2005). Thermal Regulation of Methane Hydrate Dissociation: Implications for Gas Production Models. *Energy & Fuels*, *19*(6), 2357–2363. https://doi.org/10.1021/ef0500437

Clark, P. U., Marshall, S. J., Clarke, G. K. C., Hostetler, S. W., Licciardi, J. M., & Teller, J. T. (2001). Freshwater forcing of abrupt climate change during the last glaciation. *Science*, *293*(5528), 283–287. https://doi.org/10.1126/science.1062517

Clark, Peter U., Pisias, N. G., Stocker, T. F., & Weaver, A. J. (2002). The role of the thermohaline circulation in abrupt climate change. *Nature*, Vol. 415, pp. 863–869. https://doi.org/10.1038/415863a

Clark, Peter U., Shakun, J. D., Marcott, S. A., Mix, A. C., Eby, M., Kulp, S., … Plattner, G. K. (2016). Consequences of twenty-first-century policy for multi-millennial climate and sea-level change. *Nature Climate Change*, *6*(4), 360–369. https://doi.org/10.1038/nclimate2923

Claussen, M., Brovkin, V., Ganopolski, A., Kubatzki, C., & Petoukhov, V. (2003). Climate Change in Northern Africa: The Past is Not the Future. *Climatic Change*, *57*(1), 99–118. https://doi.org/10.1023/A:1022115604225

Claypool, G. E., & Kvenvolden, K. A. (1983). Methane and other hydrocarbon gases in marine sediment. *Annual Review of Earth and Planetary Sciences. Vol. 11*, 299–327. https://doi.org/10.1146/annurev.ea.11.050183.001503

Clerc, F., Minchew, B. M., & Behn, M. D. (2019). Marine Ice Cliff Instability Mitigated by Slow Removal of Ice Shelves. *Geophysical Research Letters*, *46*(21), 12108–12116. https://doi.org/10.1029/2019GL084183

Cochrane, M. A., & Barber, C. P. (2009). Climate change, human land use and future fires in the Amazon. *Global Change Biology*, *15*(3), 601–612. https://doi.org/10.1111/j.1365-2486.2008.01786.x

Cochrane, M. A., & Schulze, M. D. (1999). Fire as a Recurrent Event in Tropical Forests of the Eastern Amazon: Effects on Forest Structure, Biomass, and Species Composition. *Biotropica*, *31*(1), 2. https://doi.org/10.2307/2663955

COHMAP Members. (1988). Climatic Changes of the Last 18,000 Years: Observations and Model Simulations. *Science*, *241*(4869), 1043 LP – 1052. https://doi.org/10.1126/science.241.4869.1043

Coles, S. L., & Brown, B. E. (2003). Coral bleaching - Capacity for acclimatization and adaptation. *Advances in Marine Biology*, *46*, 183–223. https://doi.org/10.1016/S0065-2881(03)46004-5

Collins, M., Sutherland, M., Bouwer, L., Cheong, S.-M., Frölicher, T. L., Jacot Des Combes, H., … Timmermans, M.-L. (2019). Extremes, Abrupt Changes and Managing Risks. In *IPCC Special Report on the Ocean and Cryosphere in a Changing Climate*. Retrieved from https://report.ipcc.ch/srocc/pdf/SROCC_FinalDraft_Chapter6.pdf

Cook, C. P., Van De Flierdt, T., Williams, T., Hemming, S. R., Iwai, M., Kobayashi, M., … Yamane, M. (2013). Dynamic behaviour of the East Antarctic ice sheet during Pliocene warmth. *Nature Geoscience*, *6*(9), 765–769. https://doi.org/10.1038/ngeo1889

Cook, K. H., & Vizy, E. K. (2008). Effects of twenty-first-century climate change on the Amazon rain forest. *Journal of Climate*, *21*(3), 542–560. https://doi.org/10.1175/2007JCLI1838.1

Cornford, S. L., Martin, D. F., Payne, A. J., Ng, E. G., Le Brocq, A. M., Gladstone, R. M., … Vaughan, D. G. (2015). Century-scale simulations of the response of the West Antarctic Ice Sheet to a warming climate. *The Cryosphere*, *9*(4), 1579–1600. https://doi.org/10.5194/tc-9-1579-2015

Cox, P. M., Pearson, D., Booth, B. B., Friedlingstein, P., Huntingford, C., Jones, C. D., & Luke, C. M. (2013). Sensitivity of tropical carbon to climate change constrained by carbon dioxide variability. *Nature*, *494*(7437), 341–344. https://doi.org/10.1038/nature11882

Cunningham, S. A., Kanzow, T., Rayner, D., Baringer, M. O., Johns, W. E., Marotzke, J., … Bryden, H. L. (2007). Temporal variability of the Atlantic meridional overturning circulation at 26.5°N. *Science*, *317*(5840), 935–938. https://doi.org/10.1126/science.1141304

Da Rocha, H. R., Manzi, A. O., Cabral, O. M., Miller, S. D., Goulden, M. L., Saleska, S. R., … Maia, J. F. (2009). Patterns of water and heat flux across a biome gradient from tropical forest to savanna in brazil. *Journal of Geophysical Research: Biogeosciences*, *114*(1). https://doi.org/10.1029/2007JG000640

De'Ath, G., Fabricius, K. E., Sweatman, H., & Puotinen, M. (2012). The 27-year decline of coral cover on the Great Barrier Reef and its causes. *Proceedings of the National Academy of Sciences of the United States of America*, *109*(44), 17995–17999. https://doi.org/10.1073/pnas.1208909109

De Groot, W. J., Flannigan, M. D., & Cantin, A. S. (2013). Climate change impacts on future boreal fire regimes. *Forest Ecology and Management*, *294*, 35–44. https://doi.org/10.1016/j.foreco.2012.09.027

de la Giroday, H.-M. C., Carroll, A. L., & Aukema, B. H. (2012). Breach of the northern Rocky Mountain geoclimatic barrier: initiation of range expansion by the mountain pine beetle. *Journal of Biogeography*, *39*(6), 1112–1123. https://doi.org/https://doi.org/10.1111/j.1365-2699.2011.02673.x

Dean, J. F., Middelburg, J. J., Röckmann, T., Aerts, R., Blauw, L. G., Egger, M., … Dolman, A. J. (2018). Methane Feedbacks to the Global Climate System in a Warmer World. *Reviews of Geophysics*, *56*(1), 207–250. https://doi.org/10.1002/2017RG000559

DeCarlo, T. M., Cohen, A. L., Barkley, H. C., Cobban, Q., Young, C., Shamberger, K. E., … Golbuu, Y. (2015). Coral macrobioerosion is accelerated by ocean acidification and nutrients. *Geology*, *43*(1), 7–10. https://doi.org/10.1130/G36147.1

DeCarlo, T. M., Gajdzik, L., Ellis, J., Coker, D. J., Roberts, M. B., Hammerman, N. M., … Berumen, M. L. (2020). Nutrient-supplying ocean currents modulate coral bleaching susceptibility. *Science Advances*, *6*(34), eabc5493. https://doi.org/10.1126/sciadv.abc5493

DeConto, R. M., & Pollard, D. (2016). Contribution of Antarctica to past and future sea-level rise. *Nature*, *531*(7596), 591–597. https://doi.org/10.1038/nature17145

DeConto, R. M., Pollard, D., Alley, R. B., Velicogna, I., Gasson, E., Gomez, N., … Dutton, A. (2021). The Paris Climate Agreement and future sea-level rise from Antarctica. *Nature*, *593*(7857), 83–89. https://doi.org/10.1038/s41586-021-03427-0

Defrance, D., Ramstein, G., Charbit, S., Vrac, M., Famien, A. M., Sultan, B., … Vanderlinden, J. P. (2017). Consequences of rapid ice sheet melting on the Sahelian population vulnerability. *Proceedings of the National Academy of Sciences of the United States of America*, *114*(25), 6533–6538. https://doi.org/10.1073/pnas.1619358114

Delworth, T. L., & Zeng, F. (2016). The Impact of the North Atlantic Oscillation on Climate through Its Influence on the Atlantic Meridional Overturning Circulation. *Journal of Climate*, *29*(3), 941–962. https://doi.org/10.1175/JCLI-D-15-0396.1

deMenocal, P., Ortiz, J., Guilderson, T., Adkins, J., Sarnthein, M., Baker, L., & Yarusinsky, M. (2000). Abrupt onset and termination of the African Humid Period:: rapid climate responses to gradual insolation forcing. *Quaternary Science Reviews*, *19*(1), 347–361. https://doi.org/https://doi.org/10.1016/S0277-3791(99)00081-5

Descombes, P., Wisz, M. S., Leprieur, F., Parravicini, V., Heine, C., Olsen, S. M., … Pellissier, L. (2015). Forecasted coral reef decline in marine biodiversity hotspots under climate change. *Global Change Biology*, *21*(7), 2479–2487. https://doi.org/https://doi.org/10.1111/gcb.12868

Dickens, G. R. (2011). Down the Rabbit Hole: toward appropriate discussion of methane release from gas hydrate systems during the Paleocene-Eocene thermal maximum and other past hyperthermal events. *Climate of the Past*, *7*(3), 831–846. https://doi.org/10.5194/cp-7-831-2011

Dickens, W. A., Kuhn, G., Leng, M. J., Graham, A. G. C., Dowdeswell, J. A., Meredith, M. P., … Smith, J. A. (2019). Enhanced glacial discharge from the eastern Antarctic Peninsula since the 1700s associated with a positive Southern Annular Mode. *Scientific Reports*, *9*(1), 14606. https://doi.org/10.1038/s41598-019-50897-4

Dieleman, C. M., Rogers, B. M., Potter, S., Veraverbeke, S., Johnstone, J. F., Laflamme, J., … Turetsky, M. R. (2020). Wildfire combustion and carbon stocks in the southern Canadian boreal forest: Implications for a warming world. *Global Change Biology*, *26*(11), 6062–6079. https://doi.org/https://doi.org/10.1111/gcb.15158

Dixit, V., Sherwood, S., Geoffroy, O., & Mantsis, D. (2018). The Role of Nonlinear Drying above the Boundary Layer in the Mid-Holocene African Monsoon. *Journal of Climate*, *31*(1), 233–249. https://doi.org/10.1175/JCLI-D-17-0234.1

Dolman, A. J., Shvidenko, A., Schepaschenko, D., Ciais, P., Tchebakova, N., Chen, T., … Schulze, E. D. (2012). An estimate of the terrestrial carbon budget of Russia using inventory-based, eddy covariance and inversion methods. *Biogeosciences*, *9*(12), 5323–5340. https://doi.org/10.5194/bg-9-5323-2012

Donner, S. D., Skirving, W. J., Little, C. M., Oppenheimer, M., & Hoegh-Guldberg, O. (2005). Global assessment of coral bleaching and required rates of adaptation under climate change. *Global Change Biology*, *11*(12), 2251–2265. https://doi.org/10.1111/j.1365-2486.2005.01073.x

Donovan, M. K., Adam, T. C., Shantz, A. A., Speare, K. E., Munsterman, K. S., Rice, M. M., … Burkepile, D. E. (2020). Nitrogen pollution interacts with heat stress to increase coral bleaching across the seascape. *Proceedings of the National Academy of Sciences*, *117*(10), 5351–5357. https://doi.org/10.1073/pnas.1915395117

Dubreuil, V., Debortoli, N., Funatsu, B., Nédélec, V., & Durieux, L. (2012). Impact of land-cover change in the Southern Amazonia climate: A case study for the region of Alta Floresta, Mato Grosso, Brazil. *Environmental Monitoring and Assessment*, *184*(2), 877–891. https://doi.org/10.1007/s10661-011-2006-x

Dusenge, M. E., Madhavji, S., & Way, D. A. (2020). Contrasting acclimation responses to elevated CO2 and warming between an evergreen and a deciduous boreal conifer. *Global Change Biology*, *26*(6), 3639–3657. https://doi.org/https://doi.org/10.1111/gcb.15084

Dutton, A., Carlson, A. E., Long, A. J., Milne, G. A., Clark, P. U., DeConto, R., … Raymo, M. E. (2015). Sea-level rise due to polar ice-sheet mass loss during past warm periods. *Science*, *349*(6244), aaa4019–aaa4019. https://doi.org/10.1126/science.aaa4019

Dyonisius, M. N., Petrenko, V. V., Smith, A. M., Hua, Q., Yang, B., Schmitt, J., … Weiss, R. F. (2020). Old carbon reservoirs were not important in the deglacial methane budget. *Science*, *367*(6480), 907–910. https://doi.org/10.1126/science.aax0504

Eastman, R., Warren, S. G., & Hahn, C. J. (2011). Variations in cloud cover and cloud types over the Ocean from surface observations, 1954-2008. *Journal of Climate*, *24*(22), 5914–5934. https://doi.org/10.1175/2011JCLI3972.1

Edwards, T. L., Brandon, M. A., Durand, G., Edwards, N. R., Golledge, N. R., Holden, P. B., … Wernecke, A. (2019). Revisiting Antarctic ice loss due to marine ice-cliff instability. *Nature*, *566*(7742), 58–64. https://doi.org/10.1038/s41586-019-0901-4

Edwards, T. L., Nowicki, S., Marzeion, B., Hock, R., Goelzer, H., Seroussi, H., … Zwinger, T. (2021). Projected land ice contributions to twenty-first-century sea level rise. *Nature*, *593*(7857), 74–82. https://doi.org/10.1038/s41586-021-03302-y

Ehlert, D., & Zickfeld, K. (2018). Irreversible ocean thermal expansion under carbon dioxide removal. *Earth System Dynamics*, *9*(1), 197–210. https://doi.org/10.5194/esd-9-197-2018

Eisenman, I., & Wettlaufer, J. S. (2009). Nonlinear threshold behavior during the loss of Arctic sea ice. *Proceedings of the National Academy of Sciences of the United States of America*, *106*(1), 28–32. https://doi.org/10.1073/pnas.0806887106

Eisenman, Ian. (2012). Factors controlling the bifurcation structure of sea ice retreat. *Journal of Geophysical Research Atmospheres*, *117*(1), 1–18. https://doi.org/10.1029/2011JD016164

Elliott, S., Maltrud, M., Reagan, M., Moridis, G., & Cameron-Smith, P. (2011). Marine methane cycle simulations for the period of early global warming. *Journal of Geophysical Research: Biogeosciences*, *116*(1), 1–13. https://doi.org/10.1029/2010JG001300

Eltahir, E. A. B., & Bras, R. L. (1994). Precipitation recycling in the Amazon basin. *Quarterly Journal of the Royal Meteorological Society*, *120*(518), 861–880. https://doi.org/10.1002/qj.49712051806

Emanuel, K. A., David Neelin, J., & Bretherton, C. S. (1994). On large-scale circulations in convecting atmospheres. *Quarterly Journal of the Royal Meteorological Society*, *120*(519), 1111–1143. https://doi.org/https://doi.org/10.1002/qj.49712051902

Enderlin, E. M., Howat, I. M., Jeong, S., Noh, M.-J., van Angelen, J. H., & van den Broeke, M. R. (2014). An improved mass budget for the Greenland ice sheet. *Geophysical Research Letters*, *41*(3), 866–872. https://doi.org/10.1002/2013GL059010

Espinoza, J. C., Ronchail, J., Marengo, J. A., & Segura, H. (2019). Contrasting North–South changes in Amazon wet-day and dry-day frequency and related atmospheric features (1981–2017). *Climate Dynamics*, *52*(9), 5413–5430. https://doi.org/10.1007/s00382-018-4462-2

Esquivel-Muelbert, A., Galbraith, D., Dexter, K. G., Baker, T. R., Lewis, S. L., Meir, P., … Phillips, O. L. (2017). Biogeographic distributions of neotropical trees reflect their directly measured drought tolerances. *Scientific Reports*, *7*(1), 1–11. https://doi.org/10.1038/s41598-017-08105-8

Estop-Aragonés, C., Olefeldt, D., Abbott, B. W., Chanton, J. P., Czimczik, C. I., Dean, J. F., … Anthony, K. W. (2020). Assessing the Potential for Mobilization of Old Soil Carbon After Permafrost Thaw: A Synthesis of 14C Measurements From the Northern Permafrost Region. *Global Biogeochemical Cycles*, *34*(9), e2020GB006672.

https://doi.org/https://doi.org/10.1029/2020GB006672

Euskirchen, E. S., McGuire, A. D., Kicklighter, D. W., Zhuang, Q., Clein, J. S., Dargaville, R. J., … Smith, N. V. (2006). Importance of recent shifts in soil thermal dynamics on growing season length, productivity, and carbon sequestration in terrestrial high-latitude ecosystems. *Global Change Biology*, *12*(4), 731–750. https://doi.org/10.1111/j.1365-2486.2006.01113.x

Extinction Rebellion. (2019). The Emergency - Extinction Rebellion. Retrieved December 4, 2019, from https://rebellion.earth/the-truth/the-emergency/

Ezer, T., Atkinson, L. P., Corlett, W. B., & Blanco, J. L. (2013). Gulf Stream's induced sea level rise and variability along the U.S. mid-Atlantic coast. *Journal of Geophysical Research: Oceans*, *118*(2), 685–697. https://doi.org/10.1002/jgrc.20091

Fasullo, J., & Webster, P. J. (2003). A hydrological definition of Indian Monsoon onset and withdrawal. *Journal of Climate*, *16*(19), 3200–3211. https://doi.org/10.1175/1520-0442(2003)016<3200a:AHDOIM>2.0.CO;2

Favier, L., Durand, G., Cornford, S. L., Gudmundsson, G. H., Gagliardini, O., Gillet-Chaulet, F., … Le Brocq, A. M. (2014). Retreat of Pine Island Glacier controlled by marine ice-sheet instability. *Nature Climate Change*, *4*(2), 117–121. https://doi.org/10.1038/nclimate2094

Ferrario, F., Beck, M. W., Storlazzi, C. D., Micheli, F., Shepard, C. C., & Airoldi, L. (2014). The effectiveness of coral reefs for coastal hazard risk reduction and adaptation. *Nature Communications*, *5*(May), 1–9. https://doi.org/10.1038/ncomms4794

Ferré, B., Jansson, P. G., Moser, M., Serov, P., Portnov, A., Graves, C. A., … Niemann, H. (2020). Reduced methane seepage from Arctic sediments during cold bottom-water conditions. *Nature Geoscience*, *13*(2), 144–148. https://doi.org/10.1038/s41561-019-0515-3

Fetterer, F., Knowles, K., Meier, W. N., Savoie, M., & Windnagel, A. K. (2017). *Sea Ice Index, Version 3.* https://doi.org/https://doi.org/10.7265/N5K072F8

Field, C. B., Behrenfeld, M. J., Randerson, J. T., & Falkowski, P. (1998). Primary production of the biosphere: Integrating terrestrial and oceanic components. *Science*, *281*(5374), 237–240. https://doi.org/10.1126/science.281.5374.237

Fisher, R. A., Koven, C. D., Anderegg, W. R. L., Christoffersen, B. O., Dietze, M. C., Farrior, C. E., … Moorcroft, P. R. (2018). Vegetation demographics in Earth System Models: A review of progress and priorities. *Global Change Biology*, *24*(1), 35–54. https://doi.org/https://doi.org/10.1111/gcb.13910

Flannigan, M., Stocks, B., Turetsky, M., & Wotton, M. (2009). Impacts of climate change on fire activity and fire management in the circumboreal forest. *Global Change Biology*, *15*(3), 549–560. https://doi.org/10.1111/j.1365-2486.2008.01660.x

Foley, J. A., Kutzbach, J. E., Coe, M. T., & Levis, S. (1994). Feedbacks between climate and boreal forests during the Holocene epoch. *Nature*, *371*(6492), 52–54. https://doi.org/10.1038/371052a0

Folland, C. K., Palmer, T. N., & Parker, D. E. (1986). Sahel rainfall and worldwide sea temperatures, 1901–85. *Nature*, *320*(6063), 602–607. https://doi.org/10.1038/320602a0

Fonseca, M. G., Alves, L. M., Aguiar, A. P. D., Arai, E., Anderson, L. O., Rosan, T. M., … de Aragão, L. E. O. e C. (2019).

Effects of climate and land-use change scenarios on fire probability during the 21st century in the Brazilian Amazon. *Global Change Biology*, *25*(9), 2931–2946. https://doi.org/https://doi.org/10.1111/gcb.14709

Fontaine, B., Gaetani, M., Ullmann, A., & Roucou, P. (2011). Time evolution of observed July–September sea surface temperature-Sahel climate teleconnection with removed quasi-global effect (1900–2008). *Journal of Geophysical Research: Atmospheres*, *116*(D4). https://doi.org/https://doi.org/10.1029/2010JD014843

Foster, A. C., Armstrong, A. H., Shuman, J. K., Shugart, H. H., Rogers, B. M., Mack, M. C., … Ranson, K. J. (2019). Importance of tree- and species-level interactions with wildfire, climate, and soils in interior Alaska: Implications for forest change under a warming climate. *Ecological Modelling*, *409*(March), 108765. https://doi.org/10.1016/j.ecolmodel.2019.108765

Frajka-Williams, E. (2015). Estimating the Atlantic overturning at 26°N using satellite altimetry and cable measurements. *Geophysical Research Letters*, *42*(9), 3458–3464. https://doi.org/10.1002/2015GL063220

Fraser, R. H., Kokelj, S. V., Lantz, T. C., McFarlane-Winchester, M., Olthof, I., & Lacelle, D. (2018). Climate sensitivity of high arctic permafrost terrain demonstrated by widespread ice-wedge thermokarst on banks Island. *Remote Sensing*, *10*(6). https://doi.org/10.3390/rs10060954

Fretwell, P., Pritchard, H. D., Vaughan, D. G., Bamber, J. L., Barrand, N. E., Bell, R., … Zirizzotti, A. (2013). Bedmap2: improved ice bed, surface and thickness datasets for Antarctica. *The Cryosphere*, *7*(1), 375–393. https://doi.org/10.5194/tc-7-375-2013

Frieler, K., Meinshausen, M., Golly, A., Mengel, M., Lebek, K., Donner, S. D., & Hoegh-Guldberg, O. (2013). Limiting global warming to 2C is unlikely to save most coral reefs. *Nature Climate Change*, *3*(2), 165–170. https://doi.org/10.1038/nclimate1674

Fu, R., Yin, L., Li, W., Arias, P. A., Dickinson, R. E., Huang, L., … Myneni, R. B. (2013). Increased dry-season length over southern Amazonia in recent decades and its implication for future climate projection. *Proceedings of the National Academy of Sciences*, *110*(45), 18110 LP – 18115. https://doi.org/10.1073/pnas.1302584110

Fu, Y., Li, F., Karstensen, J., & Wang, C. (2020). A stable Atlantic Meridional Overturning Circulation in a changing North Atlantic Ocean since the 1990s. *Science Advances*, *6*(48). https://doi.org/10.1126/sciadv.abc7836

Ganopolski, A., Winkelmann, R., & Schellnhuber, H. J. (2016). Critical insolation-CO2 relation for diagnosing past and future glacial inception. *Nature*, *529*(7585), 200–203. https://doi.org/10.1038/nature16494

Garbe, J., Albrecht, T., Levermann, A., Donges, J. F., & Winkelmann, R. (2020). The hysteresis of the Antarctic Ice Sheet. *Nature*, *585*(7826), 538–544. https://doi.org/10.1038/s41586-020-2727-5

Garcia-Tigreros, F., & Kessler, J. D. (2018). Limited Acute Influence of Aerobic Methane Oxidation on Ocean Carbon Dioxide and pH in Hudson Canyon, Northern U.S. Atlantic Margin. *Journal of Geophysical Research: Biogeosciences*, *123*(7), 2135–2144. https://doi.org/https://doi.org/10.1029/2018JG004384

Garcia-Tigreros, F., Leonte, M., Ruppel, C. D., Ruiz-Angulo, A., Joung, D. J., Young, B., & Kessler, J. D. (2021). Estimating the Impact of Seep Methane Oxidation on Ocean pH and Dissolved Inorganic Radiocarbon Along the U.S. Mid-Atlantic

Bight. *Journal of Geophysical Research: Biogeosciences*, *126*(1), e2019JG005621. https://doi.org/https://doi.org/10.1029/2019JG005621

Gardner, T. A., Côté, I. M., Gill, J. A., Grant, A., & Watkinson, A. R. (2003). Long-term region-wide declines in Caribbean corals. *Science*, *301*(5635), 958–960. https://doi.org/10.1126/science.1086050

Gasse, F. (2000). Hydrological changes in the African tropics since the Last Glacial Maximum. *Quaternary Science Reviews*, *19*(1), 189–211. https://doi.org/https://doi.org/10.1016/S0277-3791(99)00061-X

Gastineau, G., L'Hévéder, B., Codron, F., & Frankignoul, C. (2016). Mechanisms determining the winter atmospheric response to the Atlantic overturning circulation. *Journal of Climate*, *29*(10), 3767–3785. https://doi.org/10.1175/JCLI-D-15-0326.1

Gatti, L. V, Basso, L. S., Miller, J. B., Gloor, M., Gatti Domingues, L., Cassol, H. L. G., … Neves, R. A. L. (2021). Amazonia as a carbon source linked to deforestation and climate change. *Nature*, *595*(7867), 388–393. https://doi.org/10.1038/s41586-021-03629-6

Gauthier, S., Bernier, P., Kuuluvainen, T., Shvidenko, A. Z., & Schepaschenko, D. G. (2015, August 21). Boreal forest health and global change. *Science*, Vol. 349, pp. 819–822. https://doi.org/10.1126/science.aaa9092

Giannini, A., Saravanan, R., & Chang, P. (2003). Oceanic Forcing of Sahel Rainfall on Interannual to Interdecadal Time Scales. *Science*, *302*(5647), 1027 LP – 1030. https://doi.org/10.1126/science.1089357

Gibbs, H. K., Brown, S., Niles, J. O., & Foley, J. A. (2007). Monitoring and estimating tropical forest carbon stocks: Making REDD a reality. *Environmental Research Letters*, *2*(4). https://doi.org/10.1088/1748-9326/2/4/045023

Gibson, C. M., Brinkman, T., Cold, H., Brown, D., & Turetsky, M. (2021). Identifying increasing risks of hazards for northern land-users caused by permafrost thaw: integrating scientific and community-based research approaches. *Environmental Research Letters*, *16*(6), 64047. https://doi.org/10.1088/1748-9326/abfc79

Gibson, C. M., Chasmer, L. E., Thompson, D. K., Quinton, W. L., Flannigan, M. D., & Olefeldt, D. (2018). Wildfire as a major driver of recent permafrost thaw in boreal peatlands. *Nature Communications*, *9*(1). https://doi.org/10.1038/s41467-018-05457-1

Global Sea Level Budget Group. (2018). Global sea-level budget 1993-present. *Earth System Science Data*, *10*(3), 1551–1590. https://doi.org/10.5194/essd-10-1551-2018

Goddard, P. B., Yin, J., Griffies, S. M., & Zhang, S. (2015). An extreme event of sea-level rise along the Northeast coast of North America in 2009-2010. *Nature Communications*, *6*. https://doi.org/10.1038/ncomms7346

Goetz, S. J., MacK, M. C., Gurney, K. R., Randerson, J. T., & Houghton, R. A. (2007). Ecosystem responses to recent climate change and fire disturbance at northern high latitudes: Observations and model results contrasting northern Eurasia and North America. *Environmental Research Letters*, *2*(4). https://doi.org/10.1088/1748-9326/2/4/045031

Golledge, N. R., Kowalewski, D. E., Naish, T. R., Levy, R. H., Fogwill, C. J., & Gasson, E. G. W. (2015). The multi-millennial Antarctic commitment to future sea-level rise. *Nature*, *526*(7573), 421–425. https://doi.org/10.1038/nature15706

Golledge, Nicholas R., Keller, E. D., Gomez, N., Naughten, K. A., Bernales, J., Trusel, L. D., & Edwards, T. L. (2019). Global environmental consequences of twenty-first-century ice-sheet melt. *Nature*, *566*(7742), 65–72. https://doi.org/10.1038/s41586-019-0889-9

Gomes, V. H. F., Vieira, I. C. G., Salomão, R. P., & ter Steege, H. (2019). Amazonian tree species threatened by deforestation and climate change. *Nature Climate Change*, *9*(7), 547–553. https://doi.org/10.1038/s41558-019-0500-2

Good, P., Bamber, J., Halladay, K., Harper, A. B., Jackson, L. C., Kay, G., … Williamson, P. (2018). Recent progress in understanding climate thresholds: Ice sheets, the Atlantic meridional overturning circulation, tropical forests and responses to ocean acidification. *Progress in Physical Geography*, *42*(1), 24–60. https://doi.org/10.1177/0309133317751843

Good, P., Jones, C., Lowe, J., Betts, R., & Gedney, N. (2013). Comparing tropical forest projections from two generations of hadley centre earth system models, HadGEM2-ES and HadCM3LC. *Journal of Climate*, *26*(2), 495–511. https://doi.org/10.1175/JCLI-D-11-00366.1

Gornitz, V., & Fung, I. (1994). Potential distribution of methane hydrates in the world's oceans. *Global Biogeochemical Cycles*, *8*(3), 335–347. https://doi.org/10.1029/94GB00766

Goswami, B. N., & Xavier, P. K. (2005). ENSO control on the south Asian monsoon through the length of the rainy season. *Geophysical Research Letters*, *32*(18). https://doi.org/https://doi.org/10.1029/2005GL023216

Greenop, R., Foster, G. L., Wilson, P. A., & Lear, C. H. (2014). Middle Miocene climate instability associated with high-amplitude CO2 variability. *Paleoceanography*, *29*(9), 845–853. https://doi.org/10.1002/2014PA002653

Gregory, J M, George, S. E., & Smith, R. S. (2020). Large and irreversible future decline of the Greenland ice sheet. *The Cryosphere*, *14*(12), 4299–4322. https://doi.org/10.5194/tc-14-4299-2020

Gregory, Jonathan M., Huybrechts, P., & Raper, S. C. B. (2004). Threatened loss of the Greenland ice-sheet. *Nature*, *428*(6983), 616. https://doi.org/10.1038/428616a

Gruber, S. (2012). Derivation and analysis of a high-resolution estimate of global permafrost zonation. *The Cryosphere*, *6*(1), 221–233. https://doi.org/10.5194/tc-6-221-2012

Guest, J. R., Edmunds, P. J., Gates, R. D., Kuffner, I. B., Andersson, A. J., Barnes, B. B., … Toth, L. T. (2018). A framework for identifying and characterising coral reef "oases" against a backdrop of degradation. *Journal of Applied Ecology*, *55*(6), 2865–2875. https://doi.org/https://doi.org/10.1111/1365-2664.13179

Gupta, A., Lachance, J., Sloan, E. D., & Koh, C. A. (2008). Measurements of methane hydrate heat of dissociation using high pressure differential scanning calorimetry. *Chemical Engineering Science*, *63*(24), 5848–5853. https://doi.org/10.1016/j.ces.2008.09.002

Haarsma, R. J., Selten, F. M., & Drijfhout, S. S. (2015). Decelerating Atlantic meridional overturning circulation main cause of future west European summer atmospheric circulation changes. *Environmental Research Letters*, *10*(9). https://doi.org/10.1088/1748-9326/10/9/094007

Haine, T. W. N., & Martin, T. (2017). The Arctic-Subarctic sea ice system is entering a seasonal regime: Implications for future Arctic amplification. *Scientific Reports*, *7*(1), 1–9. https://doi.org/10.1038/s41598-017-04573-0

Halpern, B. S., Frazier, M., Potapenko, J., Casey, K. S., Koenig, K., Longo, C., … Walbridge, S. (2015). Spatial and temporal changes in cumulative human impacts on the world's ocean. *Nature Communications*, *6*(May), 1–7. https://doi.org/10.1038/ncomms8615

Halpern, D., & Woiceshyn, P. M. (1999). Onset of the Somali Jet in the Arabian Sea during June 1997. *Journal of Geophysical Research: Oceans*, *104*(C8), 18041–18046. https://doi.org/https://doi.org/10.1029/1999JC900141

Halsey, L. A., Vitt, D. H., & Zoltai, S. C. (1995). Disequilibrium response of permafrost in boreal continental western Canada to climate change. *Climatic Change*, *30*(1), 57–73. https://doi.org/10.1007/BF01093225

Hamlington, B. D., Gardner, A. S., Ivins, E., Lenaerts, J. T. M., Reager, J. T., Trossman, D. S., … Willis, M. J. (2020). Understanding of Contemporary Regional Sea-level Change and the Implications for the Future. In *Reviews of Geophysics*. https://doi.org/10.1029/2019rg000672

Hanes, C. C., Wang, X., Jain, P., Parisien, M. A., Little, J. M., & Flannigan, M. D. (2019). Fire-regime changes in canada over the last half century. *Canadian Journal of Forest Research*, *49*(3), 256–269. https://doi.org/10.1139/cjfr-2018-0293

Hansen, W. D., Fitzsimmons, R., Olnes, J., & Williams, A. P. (2020). An alternate vegetation type proves resilient and persists for decades following forest conversion in the North American boreal biome. *Journal of Ecology*, *n/a*(n/a). https://doi.org/https://doi.org/10.1111/1365-2745.13446

Harvey, L. D. D., & Huang, Z. (1995). Evaluation of the potential impact of methane clathrate destabilization on future global warming. *Journal of Geophysical Research*, *100*(D2), 2905. https://doi.org/10.1029/94JD02829

Hausfather, Z., & Peters, G. P. (2020). Emissions - the "business as usual" story is misleading. *Nature*, *577*, 618–620.

Heffernan, L., Estop-Aragonés, C., Knorr, K.-H., Talbot, J., & Olefeldt, D. (2020). Long-term Impacts of Permafrost Thaw on Carbon Storage in Peatlands: Deep Losses Offset by Surficial Accumulation. *Journal of Geophysical Research: Biogeosciences*, *125*(3), e2019JG005501. https://doi.org/https://doi.org/10.1029/2019JG005501

Heinze, C., Meyer, S., Goris, N., Anderson, L., Steinfeldt, R., Chang, N., … Bakker, D. C. E. (2015). The ocean carbon sink - Impacts, vulnerabilities and challenges. *Earth System Dynamics*, *6*(1), 327–358. https://doi.org/10.5194/esd-6-327-2015

Held, I. M., Delworth, T. L., Lu, J., Findell, K. L., & Knutson, T. R. (2005). Simulation of Sahel drought in the 20th and 21st centuries. *Proceedings of the National Academy of Sciences of the United States of America*, *102*(50), 17891–17896. https://doi.org/10.1073/pnas.0509057102

Held, Isaac M., & Soden, B. J. (2006). Robust Responses of the Hydrological Cycle to Global Warming. *Journal of Climate*, *19*(21), 5686–5699. https://doi.org/10.1175/JCLI3990.1

Heron, S. F., Maynard, J. A., Van Hooidonk, R., & Eakin, C. M. (2016). Warming Trends and Bleaching Stress of the World's Coral Reefs 1985-2012. *Scientific Reports*, *6*. https://doi.org/10.1038/srep38402

Heuzé, C. (2017). North Atlantic deep water formation and AMOC in CMIP5 models. *Ocean Science*, *13*(4), 609–622. https://doi.org/10.5194/os-13-609-2017

Hill, S. A. (2019). Theories for Past and Future Monsoon Rainfall Changes. *Current Climate Change Reports*, *5*(3), 160–171.

https://doi.org/10.1007/s40641-019-00137-8

Hirota, M., Holmgren, M., Van Nes, E. H., & Scheffer, M. (2011). Global resilience of tropical forest and savanna to critical transitions. *Science*, *334*(6053), 232–235. https://doi.org/10.1126/science.1210657

Hmiel, B., Petrenko, V. V, Dyonisius, M. N., Buizert, C., Smith, A. M., Place, P. F., … Dlugokencky, E. (2020). Preindustrial 14CH4 indicates greater anthropogenic fossil CH 4 emissions. *Nature*, *578*(May 2019), 10–14. https://doi.org/10.1038/s41586-020-1991-8

Hodgson, G. (1999). A global assessment of human effects on coral reefs. *Marine Pollution Bulletin*, *38*(5), 345–355. https://doi.org/10.1016/S0025-326X(99)00002-8

Hoegh-Guldberg, O. (2014). Coral reefs in the anthropocene: Persistence or the end of the line? *Geological Society Special Publication*, *395*(1), 167–183. https://doi.org/10.1144/SP395.17

Hoegh-Guldberg, O., Jacob, D., Taylor, M., Bindi, M., Brown, S., Camilloni, I., … Djalante, R. (2018). Impacts of 1.5°C global warming on natural and human systems. In *Global Warming of 1.5°C. An IPCC Special Report on the impacts of global warming of 1.5°C above pre-industrial levels and related global greenhouse gas emission pathways, in the context of strengthening the global response to the threat of climate change,*.

Hoegh-Guldberg, O., Jacob, D., Taylor, M., Guillén Bolaños, T., Bindi, M., Brown, S., … Zhou, G. (2019). The human imperative of stabilizing global climate change at 1.5°C. *Science*, *365*(6459). https://doi.org/10.1126/science.aaw6974

Hoegh-Guldberg, O., Mumby, P. J., Hooten, A. J., Steneck, R. S., Greenfield, P., Gomez, E., … Hatziolos, M. E. (2007). Coral Reefs Under Rapid Climate Change and Ocean Acidification. *Science*, *318*(5857).

Hoegh-Guldberg, Ove. (2014). Coral reef sustainability through adaptation: Glimmer of hope or persistent mirage? *Current Opinion in Environmental Sustainability*, *7*, 127–133. https://doi.org/10.1016/j.cosust.2014.01.005

Hoell, A., Funk, C., Barlow, M., & Shukla, S. (2016). *Recent and Possible Future Variations in the North American Monsoon BT - The Monsoons and Climate Change: Observations and Modeling* (L. M. V. de Carvalho & C. Jones, Eds.). https://doi.org/10.1007/978-3-319-21650-8_7

Holland, M. M., Bailey, D. A., & Vavrus, S. (2011). Inherent sea ice predictability in the rapidly changing Arctic environment of the Community Climate System Model, version 3. *Climate Dynamics*, *36*(7), 1239–1253. https://doi.org/10.1007/s00382-010-0792-4

Holloway, J. E., & Lewkowicz, A. G. (2020). Half a century of discontinuous permafrost persistence and degradation in western Canada. *Permafrost and Periglacial Processes*, *31*(1), 85–96. https://doi.org/10.1002/ppp.2017

Holloway, J. E., Lewkowicz, A. G., Douglas, T. A., Li, X., Turetsky, M. R., Baltzer, J. L., & Jin, H. (2020). Impact of wildfire on permafrost landscapes: A review of recent advances and future prospects. *Permafrost and Periglacial Processes*, (September 2019), 1–12. https://doi.org/10.1002/ppp.2048

Hornbach, M. J., Saffer, D. M., & Holbrook, W. S. (2004). Critically pressured free-gas reservoirs below gas-hydrate provinces. *Nature*, *427*(6970), 142–144. https://doi.org/10.1038/nature02172

Hubau, W., Lewis, S. L., Phillips, O. L., Affum-Baffoe, K., Beeckman, H., Cuní-Sanchez, A., … Zemagho, L. (2020).

Asynchronous carbon sink saturation in African and Amazonian tropical forests. *Nature*, *579*(7797), 80–87. https://doi.org/10.1038/s41586-020-2035-0

Hudson, S. R. (2011). Estimating the global radiative impact of the sea ice-albedo feedback in the Arctic. *Journal of Geophysical Research Atmospheres*, *116*(16). https://doi.org/10.1029/2011JD015804

Hugelius, G., Loisel, J., Chadburn, S., Jackson, R. B., Jones, M., MacDonald, G., … Yu, Z. (2020). Large stocks of peatland carbon and nitrogen are vulnerable to permafrost thaw. *Proceedings of the National Academy of Sciences*, *117*(34), 20438 LP – 20446. https://doi.org/10.1073/pnas.1916387117

Hughes, T. P., Anderson, K. D., Connolly, S. R., Heron, S. F., Kerry, J. T., Lough, J. M., … Wilson, S. K. (2018). Spatial and temporal patterns of mass bleaching of corals in the Anthropocene. *Science*, *359*(6371), 80–83. https://doi.org/10.1126/science.aan8048

Hughes, T. P., Kerry, J. T., Álvarez-Noriega, M., Álvarez-Romero, J. G., Anderson, K. D., Baird, A. H., … Wilson, S. K. (2017). Global warming and recurrent mass bleaching of corals. *Nature*, *543*(7645), 373–377. https://doi.org/10.1038/nature21707

Hughes, T. P., Kerry, J. T., Baird, A. H., Connolly, S. R., Chase, T. J., Dietzel, A., … Woods, R. M. (2019). Global warming impairs stock–recruitment dynamics of corals. *Nature*, *568*(7752), 387–390. https://doi.org/10.1038/s41586-019-1081-y

Hughes, T. P., Kerry, J. T., Connolly, S. R., Baird, A. H., Eakin, C. M., Heron, S. F., … Torda, G. (2019). Ecological memory modifies the cumulative impact of recurrent climate extremes. *Nature Climate Change*, *9*(1), 40–43. https://doi.org/10.1038/s41558-018-0351-2

Huntingford, C., Zelazowski, P., Galbraith, D., Mercado, L. M., Sitch, S., Fisher, R., … Cox, P. M. (2013). Simulated resilience of tropical rainforests to CO2 -induced climate change. *Nature Geoscience*, *6*(4), 268–273. https://doi.org/10.1038/ngeo1741

Huntzinger, D. N., Schaefer, K., Schwalm, C., Fisher, J. B., Hayes, D., Stofferahn, E., … Tian, H. (2020). Evaluation of simulated soil carbon dynamics in Arctic-Boreal ecosystems. *Environmental Research Letters*, *15*(2), 25005. https://doi.org/10.1088/1748-9326/ab6784

Huybrechts, P., & De Wolde, J. (1999). The dynamic response of the Greenland and Antarctic ice sheets to multiple-century climatic warming. *Journal of Climate*, *12*(8 PART 1), 2169–2188. https://doi.org/10.1175/1520-0442(1999)012<2169:tdrotg>2.0.co;2

Huybrechts, P., Letreguilly, A., & Reeh, N. (1991). The Greenland ice sheet and greenhouse warming. *Global and Planetary Change*, *3*(4), 399–412. https://doi.org/10.1016/0921-8181(91)90119-H

INPE. (2021). *TerraBrasilis*. Retrieved from http://terrabrasilis.dpi.inpe.br/

IPCC. (2013). *Climate Change 2013: The Physical Science Basis. Contribution of Working Group I to the Fifth Assessment Report of the Intergovernmental Panel on Climate Change*. Cambridge, United Kingdom and New York, NY, USA.

IPCC. (2019). Summary for policymakers. In *IPCC Special Report on the Ocean and Cryosphere in a Changing Climate*.

https://doi.org/http://www.ipcc.ch/publications_and_data/ar4/wg2/en/spm.html

Ivanov, V. Y., Hutyra, L. R., Wofsy, S. C., Munger, J. W., Saleska, S. R., De Oliveira, R. C., & De Camargo, P. B. (2012). Root niche separation can explain avoidance of seasonal drought stress and vulnerability of overstory trees to extended drought in a mature Amazonian forest. *Water Resources Research*, *48*(12), 1–21. https://doi.org/10.1029/2012WR011972

Jackson, J. B. C., Donovan, M. K., Cramer, K. L., & Lam, W. (2014). Status and Trends of Caribbean Coral Reefs: 1970-2012. *Global Coral Reef Monitoring Network, IUCN, Gland, Switzerland*.

Jackson, L. C., Kahana, R., Graham, T., Ringer, M. A., Woollings, T., Mecking, J. V., & Wood, R. A. (2015). Global and European climate impacts of a slowdown of the AMOC in a high resolution GCM. *Climate Dynamics*, *45*(11–12), 3299–3316. https://doi.org/10.1007/s00382-015-2540-2

Jackson, L C, Dubois, C., Forget, G., Haines, K., Harrison, M., Iovino, D., … Zuo, H. (2019). The Mean State and Variability of the North Atlantic Circulation: A Perspective From Ocean Reanalyses. *Journal of Geophysical Research: Oceans*, *124*(12), 9141–9170. https://doi.org/https://doi.org/10.1029/2019JC015210

Jackson, L C, & Wood, R. A. (2018). Hysteresis and Resilience of the AMOC in an Eddy-Permitting GCM. *Geophysical Research Letters*, *45*(16), 8547–8556. https://doi.org/https://doi.org/10.1029/2018GL078104

Jackson, Laura C, Peterson, K. A., Roberts, C. D., & Wood, R. A. (2016). Recent slowing of Atlantic overturning circulation as a recovery from earlier strengthening. *Nature Geoscience*, *9*(7), 518–522. https://doi.org/10.1038/ngeo2715

Jahn, A. (2018). Reduced probability of ice-free summers for 1.5 °c compared to 2 °c warming. *Nature Climate Change*, *8*(5), 409–413. https://doi.org/10.1038/s41558-018-0127-8

James, M., Lewkowicz, A. G., Smith, S. L., & Miceli, C. M. (2013). Multi-decadal degradation and persistence of permafrost in the Alaska Highway corridor, northwest Canada. *Environmental Research Letters*, *8*(4). https://doi.org/10.1088/1748-9326/8/4/045013

Jenkins, A., Dutrieux, P., Jacobs, S., Steig, E. J., Gudmundsson, G. H., Smith, J., & Heywood, K. J. (2016). Decadal ocean forcing and Antarctic ice sheet response: Lessons from the Amundsen Sea. *Oceanography*, *29*(4), 106–117. https://doi.org/10.5670/oceanog.2016.103

Jiang, L., Zhang, F., Guo, M. L., Guo, Y. J., Zhang, Y. Y., Zhou, G. W., … Huang, H. (2018). Increased temperature mitigates the effects of ocean acidification on the calcification of juvenile Pocillopora damicornis, but at a cost. *Coral Reefs*, *37*(1), 71–79. https://doi.org/10.1007/s00338-017-1634-1

Jin, C., Wang, B., & Liu, J. (2020). Future Changes and Controlling Factors of the Eight Regional Monsoons Projected by CMIP6 Models. *Journal of Climate*, *33*(21), 9307–9326. https://doi.org/10.1175/JCLI-D-20-0236.1

Jin, Q., & Wang, C. (2017). A revival of Indian summer monsoon rainfall since 2002. *Nature Climate Change*, *7*(8), 587–594. https://doi.org/10.1038/NCLIMATE3348

Jobbágy, E. G., & Jackson, R. B. (2000). The vertical distribution of soil organic carbon and its relation to climate and vegetation. *Ecological Applications*, *10*(2), 423–436. https://doi.org/10.1890/1051-

0761(2000)010[0423:TVDOSO]2.0.CO;2

Johnstone, J. F., Hollingsworth, T. N., Chapin, F. S., & Mack, M. C. (2010). Changes in fire regime break the legacy lock on successional trajectories in Alaskan boreal forest. *Global Change Biology*, *16*(4), 1281–1295. https://doi.org/10.1111/j.1365-2486.2009.02051.x

Jokiel, P. L., & Coles, S. L. (1990). Response of Hawaiian and other Indo-Pacific reef corals to elevated temperature. *Coral Reefs*, *8*(4), 155–162. https://doi.org/10.1007/BF00265006

Jones, P. D. (1994). Hemispheric surface air temperature variations: a reanalysis and an update to 1993. *Journal of Climate*, *7*(11), 1794–1802. https://doi.org/10.1175/1520-0442(1994)007<1794:HSATVA>2.0.CO;2

Jorgenson, M. T., Romanovsky, V., Harden, J., Shur, Y., Donnell, J. O., Schuur, E. A. G., & Kanevskiy, M. (2010). Resilience and vulnerability of permafrost to climate change. *Canadian Journal of Forest Research*, 1219–1236.

Joughin, I., Smith, B. E., & Medley, B. (2014). Marine ice sheet collapse potentially under way for the thwaites glacier basin, West Antarctica. *Science*, *344*(6185), 735–738. https://doi.org/10.1126/science.1249055

Ju, J., & Masek, J. G. (2016). The vegetation greenness trend in Canada and US Alaska from 1984-2012 Landsat data. *Remote Sensing of Environment*, *176*, 1–16. https://doi.org/10.1016/j.rse.2016.01.001

Jury, C. P., & Toonen, R. J. (2019). Adaptive responses and local stressor mitigation drive coral resilience in warmer, more acidic oceans. *Proceedings of the Royal Society B: Biological Sciences*, *286*(1902). https://doi.org/10.1098/rspb.2019.0614

Karspeck, A. R., Stammer, D., Köhl, A., Danabasoglu, G., Balmaseda, M., Smith, D. M., … Rosati, A. (2017). Comparison of the Atlantic meridional overturning circulation between 1960 and 2007 in six ocean reanalysis products. *Climate Dynamics*, *49*(3), 957–982. https://doi.org/10.1007/s00382-015-2787-7

Kasischke, E. S. (2000). Boreal Ecosystems in the Global Carbon Cycle. In E. S. Kasischke & B. J. Stocks (Eds.), *Fire, Climate Change, and Carbon Cycling in the Boreal Forest* (pp. 19–30). https://doi.org/10.1007/978-0-387-21629-4_2

Kasischke, E. S., & Turetsky, M. R. (2006). Recent changes in the fire regime across the North American boreal region - Spatial and temporal patterns of burning across Canada and Alaska. *Geophysical Research Letters*, *33*(9), 1–5. https://doi.org/10.1029/2006GL025677

Kay, J. E., L'Ecuyer, T., Gettelman, A., Stephens, G., & O'Dell, C. (2008). The contribution of cloud and radiation anomalies to the 2007 Arctic sea ice extent minimum. *Geophysical Research Letters*, *35*(8), 1–5. https://doi.org/10.1029/2008GL033451

Kersalé, M., Lamont, T., Speich, S., Terre, T., Laxenaire, R., Roberts, M. J., … Ansorge, I. J. (2018). Moored observations of mesoscale features in the Cape Basin: characteristics and local impacts on water mass distributions. *Ocean Science*, *14*(5), 923–945. https://doi.org/10.5194/os-14-923-2018

Kessler, J D, Reeburgh, W. S., Southon, J., & Varela, R. (2005). Fossil methane source dominates Cariaco Basin water column methane geochemistry. *Geophysical Research Letters*, *32*(12). https://doi.org/https://doi.org/10.1029/2005GL022984

Kessler, John D., Valentine, D. L., Redmond, M. C., Du, M., Chan, E. W., Mendes, S. D., … Weber, T. C. (2011). A persistent oxygen anomaly reveals the fate of spilled methane in the deep Gulf of Mexico. *Science*, *331*(6015), 312–315. https://doi.org/10.1126/science.1199697

Keuper, F., Wild, B., Kummu, M., Beer, C., Blume-Werry, G., Fontaine, S., … Dorrepaal, E. (2020). Carbon loss from northern circumpolar permafrost soils amplified by rhizosphere priming. *Nature Geoscience*, *13*(8), 560–565. https://doi.org/10.1038/s41561-020-0607-0

Kharuk, V. I., Im, S. T., & Soldatov, V. V. (2020). Siberian silkmoth outbreaks surpassed geoclimatic barrier in Siberian Mountains. *Journal of Mountain Science*, *17*(8), 1891–1900. https://doi.org/10.1007/s11629-020-5989-3

Kharuk, V. I., Ranson, K. J., & Dvinskaya, M. L. (2008). Wildfires dynamic in the larch dominance zone. *Geophysical Research Letters*, *35*(1), 1–6. https://doi.org/10.1029/2007GL032291

Khazendar, A., Rignot, E., Schroeder, D. M., Seroussi, H., Schodlok, M. P., Scheuchl, B., … Velicogna, I. (2016). Rapid submarine ice melting in the grounding zones of ice shelves in West Antarctica. *Nature Communications*, *7*, 1–8. https://doi.org/10.1038/ncomms13243

Kirschke, S., Bousquet, P., Ciais, P., Saunois, M., Canadell, J. G., Dlugokencky, E. J., … Zeng, G. (2013). Three decades of global methane sources and sinks. *Nature Geoscience*, *6*(10), 813–823. https://doi.org/10.1038/ngeo1955

Kloster, S., Mahowald, N. M., Randerson, J. T., & Lawrence, P. J. (2012). The impacts of climate, land use, and demography on fires during the 21st century simulated by CLM-CN. *Biogeosciences*, *9*(1), 509–525. https://doi.org/10.5194/bg-9-509-2012

Koenig, S. J., DeConto, R. M., & Pollard, D. (2014). Impact of reduced Arctic sea ice on Greenland ice sheet variability in a warmer than present climate. *Geophysical Research Letters*, *41*(11), 3933–3942. https://doi.org/10.1002/2014GL059770

Kokelj, S. V., Tunnicliffe, J., Lacelle, D., Lantz, T. C., Chin, K. S., & Fraser, R. (2015). Increased precipitation drives mega slump development and destabilization of ice-rich permafrost terrain, northwestern Canada. *Global and Planetary Change*, *129*, 56–68. https://doi.org/10.1016/j.gloplacha.2015.02.008

Kominz, M. A., Browning, J. V., Miller, K. G., Sugarman, P. J., Mizintseva, S., & Scotese, C. R. (2008). Late Cretaceous to Miocene sea-level estimates from the New Jersey and Delaware coastal plain coreholes: An error analysis. *Basin Research*, *20*(2), 211–226. https://doi.org/10.1111/j.1365-2117.2008.00354.x

Kopp, R. E., DeConto, R. M., Bader, D. A., Hay, C. C., Horton, R. M., Kulp, S., … Strauss, B. H. (2017). Evolving Understanding of Antarctic Ice-Sheet Physics and Ambiguity in Probabilistic Sea-Level Projections. *Earth's Future*, *5*(12), 1217–1233. https://doi.org/10.1002/2017EF000663

Kopp, R. E., Shwom, R. L., Wagner, G., & Yuan, J. (2016, August 1). Tipping elements and climate–economic shocks: Pathways toward integrated assessment. *Earth's Future*, Vol. 4, pp. 346–372. https://doi.org/10.1002/2016EF000362

Koven, C D, Schuur, E. A. G., Schädel, C., Bohn, T. J., Burke, E. J., Chen, G., … Turetsky, M. (2015). A simplified, data-constrained approach to estimate the permafrost carbon–climate feedback. *Philosophical Transactions of the Royal Society A: Mathematical, Physical and Engineering Sciences*, *373*(2054), 20140423. https://doi.org/10.1098/rsta.2014.0423

Koven, Charles D., Lawrence, D. M., & Riley, W. J. (2015). Permafrost carbon-climate feedback is sensitive to deep soil carbon decomposability but not deep soil nitrogen dynamics. *Proceedings of the National Academy of Sciences of the United States of America*, *112*(12), 3752–3757. https://doi.org/10.1073/pnas.1415123112

Koven, Charles D., Riley, W. J., & Stern, A. (2013). Analysis of permafrost thermal dynamics and response to climate change in the CMIP5 earth system models. *Journal of Climate*, *26*(6), 1877–1900. https://doi.org/10.1175/JCLI-D-12-00228.1

Kretschmer, K., Biastoch, A., Rüpke, L., & Burwicz, E. (2015). Modeling the fate of methane hydrates under global warming. *Global Biogeochemical Cycles*, *29*(5), 610–625. https://doi.org/10.1002/2014GB005011

Kretschmer, M., Zappa, G., & Shepherd, T. G. (2020). The role of Barents--Kara sea ice loss in projected polar vortex changes. *Weather and Climate Dynamics*, *1*(2), 715–730. https://doi.org/10.5194/wcd-1-715-2020

Kriegler, E., Hall, J. W., Held, H., Dawson, R., & Schellnhuber, H. J. (2009). Imprecise probability assessment of tipping points in the climate system. *Proceedings of the National Academy of Sciences of the United States of America*, *106*(13), 5041–5046. https://doi.org/10.1073/pnas.0809117106

Krishnamurti, T. N., Ardanuy, P., Ramanathan, Y., & Pasch, R. (1981). On the Onset Vortex of the Summer Monsoon. *Monthly Weather Review*, *109*(2), 344–363. https://doi.org/10.1175/1520-0493(1981)109<0344:OTOVOT>2.0.CO;2

Kroeker, K. J., Kordas, R. L., Crim, R. N., & Singh, G. G. (2010). Meta-analysis reveals negative yet variable effects of ocean acidification on marine organisms. *Ecology Letters*, *13*(11), 1419–1434. https://doi.org/10.1111/j.1461-0248.2010.01518.x

Kurz, W. A., Stinson, G., Rampley, G. J., Dymond, C. C., & Neilson, E. T. (2008). Risk of natural disturbances makes future contribution of Canada's forests to the global carbon cycle highly uncertain. *Proceedings of the National Academy of Sciences of the United States of America*, *105*(5), 1551–1555. https://doi.org/10.1073/pnas.0708133105

Kvenvolden, K. A. (1988). Methane hydrate - A major reservoir of carbon in the shallow geosphere? *Chemical Geology*, *71*(1–3), 41–51. https://doi.org/10.1016/0009-2541(88)90104-0

Kwok, R., & Rothrock, D. A. (2009). Decline in Arctic sea ice thickness from submarine and ICESat records: 1958-2008. *Geophysical Research Letters*, *36*(15), 1–5. https://doi.org/10.1029/2009GL039035

Lammertsma, E. I., De Boer, H. J., Dekker, S. C., Dilcher, D. L., Lotter, A. F., & Wagner-Cremer, F. (2011). Global CO2 rise leads to reduced maximum stomatal conductance in Florida vegetation. *Proceedings of the National Academy of Sciences of the United States of America*, *108*(10), 4035–4040. https://doi.org/10.1073/pnas.1100371108

Lapola, D. M., Oyama, M. D., & Nobre, C. A. (2009). Exploring the range of climate biome projections for tropical South America: The role of CO2 fertilization and seasonality. *Global Biogeochemical Cycles*, *23*(3). https://doi.org/https://doi.org/10.1029/2008GB003357

Lara, M. J., McGuire, A. D., Euskirchen, E. S., Genet, H., Yi, S., Rutter, R., … Wullschleger, S. D. (2020). Local-scale Arctic tundra heterogeneity affects regional-scale carbon dynamics. *Nature Communications*, *11*(1), 4925. https://doi.org/10.1038/s41467-020-18768-z

Laufkötter, C., Zscheischler, J., & Frölicher, T. L. (2020). High-impact marine heatwaves attributable to human-induced global warming. *Science*, *369*(6511), 1621 LP – 1625. https://doi.org/10.1126/science.aba0690

Laurion, I., Massicotte, P., Mazoyer, F., Negandhi, K., & Mladenov, N. (2020). Weak mineralization despite strong processing of dissolved organic matter in Eastern Arctic tundra ponds. *Limnology and Oceanography*, *n/a*(n/a). https://doi.org/https://doi.org/10.1002/lno.11634

Lauvset, S. K., Gruber, N., Landschützer, P., Olsen, A., & Tjiputra, J. (2015). Trends and drivers in global surface ocean pH over the past 3 decades. *Biogeosciences*, *12*(5), 1285–1298. https://doi.org/10.5194/bg-12-1285-2015

Lawrence, D. M., Slater, A. G., & Swenson, S. C. (2012). Simulation of Present-Day and Future Permafrost and Seasonally Frozen Ground Conditions in CCSM4. *Journal of Climate*, *25*(7), 2207–2225. https://doi.org/10.1175/JCLI-D-11-00334.1

Lawrence, D., & Vandecar, K. (2015). Effects of tropical deforestation on climate and agriculture. *Nature Climate Change*, *5*(1), 27–36. https://doi.org/10.1038/nclimate2430

Le Brocq, A. M., Payne, A. J., & Vieli, A. (2010). An improved Antarctic dataset for high resolution numerical ice sheet models (ALBMAP v1). *Earth System Science Data*, *2*(2), 247–260. https://doi.org/10.5194/essd-2-247-2010

Ledru, M. P., Salgado-Labouriau, M. L., & Lorscheitter, M. L. (1998). Vegetation dynamics in southern and central Brazil during the last 10,000 yr B.P. *Review of Palaeobotany and Palynology*, *99*(2), 131–142. https://doi.org/10.1016/S0034-6667(97)00049-3

Lee, H., Swenson, S. C., Slater, A. G., & Lawrence, D. M. (2014). Effects of excess ground ice on projections of permafrost in a warming climate. *Environmental Research Letters*, *9*(12). https://doi.org/10.1088/1748-9326/9/12/124006

Lee, J.-E., Risi, C., Fung, I., Worden, J., Scheepmaker, R. A., Lintner, B., & Frankenberg, C. (2012). Asian monsoon hydrometeorology from TES and SCIAMACHY water vapor isotope measurements and LMDZ simulations: Implications for speleothem climate record interpretation. *Journal of Geophysical Research: Atmospheres*, *117*(D15). https://doi.org/https://doi.org/10.1029/2011JD017133

Leite-Filho, A. T., de Sousa Pontes, V. Y., & Costa, M. H. (2019). Effects of Deforestation on the Onset of the Rainy Season and the Duration of Dry Spells in Southern Amazonia. *Journal of Geophysical Research: Atmospheres*, *124*(10), 5268–5281. https://doi.org/https://doi.org/10.1029/2018JD029537

Leite-Filho, A. T., Soares-Filho, B. S., Davis, J. L., Abrahão, G. M., & Börner, J. (2021). Deforestation reduces rainfall and agricultural revenues in the Brazilian Amazon. *Nature Communications*, *12*(1), 2591. https://doi.org/10.1038/s41467-021-22840-7

Lenaerts, J. T. M., Van Angelen, J. H., Van Den Broeke, M. R., Gardner, A. S., Wouters, B., & Van Meijgaard, E. (2013). Irreversible mass loss of Canadian Arctic Archipelago glaciers. *Geophysical Research Letters*, *40*(5), 870–874. https://doi.org/10.1002/grl.50214

Lenton, T. M. (2011, July). Early warning of climate tipping points. *Nature Climate Change*, Vol. 1, pp. 201–209. https://doi.org/10.1038/nclimate1143

Lenton, T. M. (2012, February). Arctic climate tipping points. *Ambio*, Vol. 41, pp. 10–22. https://doi.org/10.1007/s13280-011-

0221-x

Lenton, T. M., Held, H., Kriegler, E., Hall, J. W., Lucht, W., Rahmstorf, S., & Schellnhuber, H. J. (2008, February 12). Tipping elements in the Earth's climate system. *Proceedings of the National Academy of Sciences of the United States of America*, Vol. 105, pp. 1786–1793. https://doi.org/10.1073/pnas.0705414105

Lenton, T. M., Rockström, J., Gaffney, O., Rahmstorf, S., Richardson, K., Steffen, W., & Schellnhuber, H. J. (2019). Climate tipping points — too risky to bet against. *Nature*, *575*(7784), 592–595. https://doi.org/10.1038/d41586-019-03595-0

Lenton, T. M., & Schellnhuber, H. J. (2007). Tipping the scales. *Nature Climate Change*, Vol. 1, pp. 97–98.

Leonte, M., Ruppel, C. D., Ruiz-Angulo, A., & Kessler, J. D. (2020). Surface Methane Concentrations Along the Mid-Atlantic Bight Driven by Aerobic Subsurface Production Rather Than Seafloor Gas Seeps. *Journal of Geophysical Research: Oceans*, *125*(5), e2019JC015989. https://doi.org/https://doi.org/10.1029/2019JC015989

Levermann, A., Griesel, A., Hofmann, M., Montoya, M., & Rahmstorf, S. (2005). Dynamic sea level changes following changes in the thermohaline circulation. *Climate Dynamics*, *24*(4), 347–354. https://doi.org/10.1007/s00382-004-0505-y

Levermann, A., Petoukhov, V., Schewe, J., & Schellnhuber, H. J. (2016). Abrupt monsoon transitions as seen in paleorecords can be explained by moisture-advection feedback. *Proceedings of the National Academy of Sciences*, *113*(17), E2348--E2349. https://doi.org/10.1073/pnas.1603130113

Levermann, A., Schewe, J., Petoukhov, V., & Held, H. (2009). Basic mechanism for abrupt monsoon transitions. *Proceedings of the National Academy of Sciences of the United States of America*, *106*(49), 20572–20577. https://doi.org/10.1073/pnas.0901414106

Lewkowicz, A. G., & Way, R. G. (2019). Extremes of summer climate trigger thousands of thermokarst landslides in a High Arctic environment. *Nature Communications*, *10*(1), 1–11. https://doi.org/10.1038/s41467-019-09314-7

Lhermitte, S., Sun, S., Shuman, C., Wouters, B., Pattyn, F., Wuite, J., … Nagler, T. (2020). Damage accelerates ice shelf instability and mass loss in Amundsen Sea Embayment. *Proceedings of the National Academy of Sciences*, *117*(40), 24735 LP – 24741. https://doi.org/10.1073/pnas.1912890117

Li, C., Notz, D., Tietsche, S., & Marotzke, J. (2013). The transient versus the equilibrium response of sea ice to global warming. *Journal of Climate*, *26*(15), 5624–5636. https://doi.org/10.1175/JCLI-D-12-00492.1

Li, F., Lozier, M. S., Bacon, S., Bower, A. S., Cunningham, S. A., de Jong, M. F., … Zhou, C. (2021). Subpolar North Atlantic western boundary density anomalies and the Meridional Overturning Circulation. *Nature Communications*, *12*(1), 3002. https://doi.org/10.1038/s41467-021-23350-2

Li, W., & Fu, R. (2004). Transition of the large-scale atmospheric and land surface conditions from the dry to the wet season over Amazonia as diagnosed by the ECMWF re-analysis. *Journal of Climate*, *17*(13), 2637–2651. https://doi.org/10.1175/1520-0442(2004)017<2637:TOTLAA>2.0.CO;2

Li, W., Fu, R., & Dickinson, R. E. (2006). Rainfall and its seasonality over the Amazon in the 21st century as assessed by the coupled models for the IPCC AR4. *Journal of Geophysical Research Atmospheres*, *111*(2), 1–14.

https://doi.org/10.1029/2005JD006355

Li, W., Fu, R., Juárez, R. I. N., & Fernandes, K. (2008). Observed change of the standardized precipitation index, its potential cause and implications to future climate change in the Amazon region. *Philosophical Transactions of the Royal Society B: Biological Sciences*, *363*(1498), 1767–1772. https://doi.org/10.1098/rstb.2007.0022

Liang, X.-Z., & Wang, W.-C. (1998). Associations between China monsoon rainfall and tropospheric jets. *Quarterly Journal of the Royal Meteorological Society*, *124*(552), 2597–2623. https://doi.org/https://doi.org/10.1002/qj.49712455204

Lin, J. L., Qian, T., & Shinoda, T. (2014). Stratocumulus clouds in Southeastern pacific simulated by eight CMIP5-CFMIP global climate models. *Journal of Climate*, *27*(8), 3000–3022. https://doi.org/10.1175/JCLI-D-13-00376.1

Lin, X., Rogers, B. M., Sweeney, C., Chevallier, F., Arshinov, M., Dlugokencky, E., … Keppel-Aleks, G. (2020). Siberian and temperate ecosystems shape Northern Hemisphere atmospheric CO2 seasonal amplification. *Proceedings of the National Academy of Sciences*, *117*(35), 21079–21087. https://doi.org/10.1073/pnas.1914135117

Little, C. M., Hu, A., Hughes, C. W., McCarthy, G. D., Piecuch, C. G., Ponte, R. M., & Thomas, M. D. (2019). The Relationship Between U.S. East Coast Sea Level and the Atlantic Meridional Overturning Circulation: A Review. *Journal of Geophysical Research: Oceans*, *124*(9), 6435–6458. https://doi.org/https://doi.org/10.1029/2019JC015152

Liu, W., Fedorov, A. V, Xie, S.-P., & Hu, S. (2020). Climate impacts of a weakened Atlantic Meridional Overturning Circulation in a warming climate. *Science Advances*, *6*(26), eaaz4876. https://doi.org/10.1126/sciadv.aaz4876

Liu, W., & Liu, Z. (2013). A diagnostic indicator of the stability of the atlantic meridional overturning circulation in CCSM3. *Journal of Climate*, *26*(6), 1926–1938. https://doi.org/10.1175/JCLI-D-11-00681.1

Liu, W., Liu, Z., & Brady, E. C. (2014). Why is the AMOC monostable in coupled general circulation models? *Journal of Climate*, *27*(6), 2427–2443. https://doi.org/10.1175/JCLI-D-13-00264.1

Liu, W., Xie, S.-P., Liu, Z., & Zhu, J. (2017). Overlooked possibility of a collapsed Atlantic Meridional Overturning Circulation in warming climate. *Science Advances*, *3*(1), e1601666. https://doi.org/10.1126/sciadv.1601666

Liu, Y., Moore, J. C., Cheng, X., Gladstone, R. M., Bassis, J. N., Liu, H., … Hui, F. (2015). Ocean-driven thinning enhances iceberg calving and retreat of Antarctic ice shelves. *Proceedings of the National Academy of Sciences of the United States of America*, *112*(11), 3263–3268. https://doi.org/10.1073/pnas.1415137112

Liu, Z., Ballantyne, A. P., & Cooper, L. A. (2019). Biophysical feedback of global forest fires on surface temperature. *Nature Communications*, *10*(1), 1–9. https://doi.org/10.1038/s41467-018-08237-z

Liu, Z., Kimball, J. S., Parazoo, N. C., Ballantyne, A. P., Wang, W. J., Madani, N., … Euskirchen, E. S. (2020). Increased high-latitude photosynthetic carbon gain offset by respiration carbon loss during an anomalous warm winter to spring transition. *Global Change Biology*, *26*(2), 682–696. https://doi.org/https://doi.org/10.1111/gcb.14863

Lovejoy, T. E., & Nobre, C. (2018, February 21). Amazon tipping point. *Science Advances*, Vol. 4. https://doi.org/10.1126/sciadv.aat2340

Lovejoy, T. E., & Nobre, C. (2019). Amazon tipping point: Last chance for action. *Science Advances*, *5*(12), 4–6. https://doi.org/10.1126/sciadv.aba2949

Lozier, M., Bacon, S., Bower, A. S., Cunningham, S. A., Femke de Jong, M., de Steur, L., … Zika, J. D. (2017). Overturning in the Subpolar North Atlantic Program: A New International Ocean Observing System. *Bulletin of the American Meteorological Society*, *98*(4), 737–752. https://doi.org/10.1175/BAMS-D-16-0057.1

Lozier, M. S., Li, F., Bacon, S., Bahr, F., Bower, A. S., Cunningham, S. A., … Zhao, J. (2019). A sea change in our view of overturning in the subpolar North Atlantic. *Science*, *363*(6426), 516–521. https://doi.org/10.1126/science.aau6592

Lucht, W., Schaphoff, S., Erbrecht, T., Heyder, U., & Cramer, W. (2006). Terrestrial vegetation redistribution and carbon balance under climate change. *Carbon Balance and Management*, *1*(1), 1–7. https://doi.org/10.1186/1750-0680-1-6

Lynch-Stieglitz, J. (2017). The Atlantic Meridional Overturning Circulation and Abrupt Climate Change. *Annual Review of Marine Science*, *9*(1), 83–104. https://doi.org/10.1146/annurev-marine-010816-060415

Lyra, A. de A., Chou, S. C., & Sampaio, G. de O. (2016). Sensitivity of the Amazon biome to high resolution climate change projections. *Acta Amazonica*, *46*(2), 175–188. https://doi.org/10.1590/1809-4392201502225

MacFerrin, M., Machguth, H., As, D. van, Charalampidis, C., Stevens, C. M., Heilig, A., … Abdalati, W. (2019). Rapid expansion of Greenland's low-permeability ice slabs. *Nature*, *573*(7774), 403–407. https://doi.org/10.1038/s41586-019-1550-3

Mack, M. C., Bret-Harte, M. S., Hollingsworth, T. N., Jandt, R. R., Schuur, E. A. G., Shaver, G. R., & Verbyla, D. L. (2011). Carbon loss from an unprecedented Arctic tundra wildfire. *Nature*, *475*(7357), 489–492. https://doi.org/10.1038/nature10283

Mack, M. C., Walker, X. J., Johnstone, J. F., Alexander, H. D., Melvin, A. M., Jean, M., & Miller, S. N. (2021). Carbon loss from boreal forest wildfires offset by increased dominance of deciduous trees. *Science*, *372*(6539), 280–283. https://doi.org/10.1126/science.abf3903

Maher, B. A. (2008). Holocene variability of the East Asian summer monsoon from Chinese cave records: a re-assessment. *The Holocene*, *18*(6), 861–866. https://doi.org/10.1177/0959683608095569

Malhi, Y., Aragão, L. E. O. C., Galbraith, D., Huntingford, C., Fisher, R., Zelazowski, P., … Meir, P. (2009). Exploring the likelihood and mechanism of a climate-change-induced dieback of the Amazon rainforest. *Proceedings of the National Academy of Sciences of the United States of America*, *106*(49), 20610–20615. https://doi.org/10.1073/pnas.0804619106

Malhi, Y., Wood, D., Baker, T. R., Wright, J., Phillips, O. L., Cochrane, T., … Vinceti, B. (2006). The regional variation of aboveground live biomass in old-growth Amazonian forests. *Global Change Biology*, *12*(7), 1107–1138. https://doi.org/10.1111/j.1365-2486.2006.01120.x

Mann, D. H., Rupp, T. S., Olson, M. A., & Duffy, P. A. (2012). Is Alaska's Boreal Forest Now Crossing a Major Ecological Threshold? *Arctic, Antarctic, and Alpine Research*, *44*(3), 319–331. https://doi.org/10.1657/1938-4246-44.3.319

Marchal, J., Cumming, S. G., & McIntire, E. J. B. (2020). Turning Down the Heat: Vegetation Feedbacks Limit Fire Regime Responses to Global Warming. *Ecosystems*, *23*(1), 204–216. https://doi.org/10.1007/s10021-019-00398-2

Marengo, J. A., Nobre, C. A., Sampaio, G., Salazar, L. F., & Borma, L. S. (2011). Climate change in the Amazon Basin: Tipping points, changes in extremes, and impacts on natural and human systems. In *Tropical Rainforest Responses to*

Climatic Change (pp. 259–283). https://doi.org/10.1007/978-3-642-05383-2_9

Martin-Español, A., Bamber, J. L., & Zammit-Mangion, A. (2017). Constraining the mass balance of East Antarctica. *Geophysical Research Letters*, *44*(9), 4168–4175. https://doi.org/10.1002/2017GL072937

Martin-StPaul, N., Delzon, S., & Cochard, H. (2017). Plant resistance to drought depends on timely stomatal closure. *Ecology Letters*, *20*(11), 1437–1447. https://doi.org/10.1111/ele.12851

Masson-Delmotte, V., Schulz, M., Abe-Ouchi, A., Beer, J., Ganopolski, A., González Rouco, J. F., … Timmermann, A. (2013). Information from paleoclimate archives. In *Climate Change 2013: The Physical Science Basis. Contribution of Working Group I to the Fifth Assessment Report of the Intergovernmental Panel on Climate Change*. https://doi.org/10.1017/CBO9781107415324.013

Massonnet, F., Fichefet, T., Goosse, H., Bitz, C. M., Philippon-Berthier, G., Holland, M. M., & Barriat, P. Y. (2012). Constraining projections of summer Arctic sea ice. *Cryosphere*, *6*(6), 1383–1394. https://doi.org/10.5194/tc-6-1383-2012

Mau, S., Blees, J., Helmke, E., Niemann, H., & Damm, E. (2013). Vertical distribution of methane oxidation and methanotrophic response to elevated methane concentrations in stratified waters of the Arctic fjord Storfjorden (Svalbard, Norway). *Biogeosciences*, *10*(10), 6267–6268. https://doi.org/10.5194/bg-10-6267-2013

McCarty, J. L., Smith, T. E. L., & Turetsky, M. R. (2020). Arctic fires re-emerging. *Nature Geoscience*, *13*(10), 658–660. https://doi.org/10.1038/s41561-020-00645-5

McGee, D., deMenocal, P. B., Winckler, G., Stuut, J. B. W., & Bradtmiller, L. I. (2013). The magnitude, timing and abruptness of changes in North African dust deposition over the last 20,000yr. *Earth and Planetary Science Letters*, *371–372*, 163–176. https://doi.org/https://doi.org/10.1016/j.epsl.2013.03.054

McGinnis, D. F., Greinert, J., Artemov, Y., Beaubien, S. E., & Wüest, A. (2006). Fate of rising methane bubbles in stratified waters: How much methane reaches the atmosphere? *Journal of Geophysical Research: Oceans*, *111*(9), 1–15. https://doi.org/10.1029/2005JC003183

Mcguire, A. D., Anderson, L. G., Christensen, T. R., Scott, D., Laodong, G., Hayes, D. J., … Nigel, R. (2009). Sensitivity of the carbon cycle in the Arctic to climate change. *Ecological Monographs*, *79*(4), 523–555. https://doi.org/10.1890/08-2025.1

McGuire, A. D., Lawrence, D. M., Koven, C., Clein, J. S., Burke, E., Chen, G., … Zhuang, Q. (2018). Dependence of the evolution of carbon dynamics in the northern permafrost region on the trajectory of climate change. *Proceedings of the National Academy of Sciences of the United States of America*, *115*(15), 3882–3887. https://doi.org/10.1073/pnas.1719903115

McInerney, F. A., & Wing, S. L. (2011). The Paleocene-Eocene Thermal Maximum: A Perturbation of Carbon Cycle, Climate, and Biosphere with Implications for the Future. *Annual Review of Earth and Planetary Sciences*, *39*(1), 489–516. https://doi.org/10.1146/annurev-earth-040610-133431

McKibben, B. (2019). The climate science is clear: it's now or never to avert catastrophe. *The Guardian*. Retrieved from https://www.theguardian.com/commentisfree/2019/nov/20/climate-crisis-its-now-or-never-to-avert-catastrophe

McMillan, M., Leeson, A., Shepherd, A., Briggs, K., Armitage, T. W. K., Hogg, A., … Gilbert, L. (2016). A high-resolution record of Greenland mass balance. *Geophysical Research Letters*, *43*(13), 7002–7010. https://doi.org/10.1002/2016GL069666

Mekonnen, Z. A., Riley, W. J., Randerson, J. T., Grant, R. F., & Rogers, B. M. (2019). Expansion of high-latitude deciduous forests driven by interactions between climate warming and fire. *Nature Plants*, *5*(9), 952–958. https://doi.org/10.1038/s41477-019-0495-8

Menary, M. B., & Wood, R. A. (2018). An anatomy of the projected North Atlantic warming hole in CMIP5 models. *Climate Dynamics*, *50*(7–8), 3063–3080. https://doi.org/10.1007/s00382-017-3793-8

Mengel, M., & Levermann, A. (2014). Ice plug prevents irreversible discharge from east Antarctica. *Nature Climate Change*, *4*(6), 451–455. https://doi.org/10.1038/nclimate2226

Menviel, L., Timmermann, A., Mouchet, A., & Timm, O. (2008). Meridional reorganizations of marine and terrestrial productivity during Heinrich events. *Paleoceanography*, *23*(1). https://doi.org/https://doi.org/10.1029/2007PA001445

Meredith, M., Sommerkorn, M., Cassotta, S., Derksen, C., Ekaykin, A., Hollowed, A., … Schuur, E. (2019). Polar regions. In *IPCC Special Report on the Ocean and Cryosphere in a Changing Climate*. https://doi.org/10.1016/S1366-7017(01)00066-6

Mestdagh, T., Poort, J., & De Batist, M. (2017, June 1). The sensitivity of gas hydrate reservoirs to climate change: Perspectives from a new combined model for permafrost-related and marine settings. *Earth-Science Reviews*, Vol. 169, pp. 104–131. https://doi.org/10.1016/j.earscirev.2017.04.013

Mies, M., Francini-Filho, R. B., Zilberberg, C., Garrido, A. G., Longo, G. O., Laurentino, E., … Banha, T. N. S. (2020). South Atlantic Coral Reefs Are Major Global Warming Refugia and Less Susceptible to Bleaching. *Frontiers in Marine Science*, *7*, 514. https://doi.org/10.3389/fmars.2020.00514

Milkov, A. V. (2004). Global estimates of hydrate-bound gas in marine sediments: How much is really out there? *Earth-Science Reviews*, *66*(3–4), 183–197. https://doi.org/10.1016/j.earscirev.2003.11.002

Millar, J. R., Nicholls, Z. R., Friedlingstein, P., & Allen, M. R. (2017). A modified impulse-response representation of the global near-surface air temperature and atmospheric concentration response to carbon dioxide emissions. *Atmospheric Chemistry and Physics*, *17*(11), 7213–7228. https://doi.org/10.5194/acp-17-7213-2017

Moat, B. I., Smeed, D. A., Frajka-Williams, E., Desbruyères, D. G., Beaulieu, C., Johns, W. E., … Bryden, H. L. (2020). Pending recovery in the strength of the meridional overturning circulation at 26{\degree}\,N. *Ocean Science*, *16*(4), 863–874. https://doi.org/10.5194/os-16-863-2020

Mollica, N. R., Guo, W., Cohen, A. L., Huang, K. F., Foster, G. L., Donald, H. K., & Solow, A. R. (2018). Ocean acidification affects coral growth by reducing skeletal density. *Proceedings of the National Academy of Sciences of the United States of America*, *115*(8), 1754–1759. https://doi.org/10.1073/pnas.1712806115

Molnar, P., Boos, W. R., & Battisti, D. S. (2010). Orographic Controls on Climate and Paleoclimate of Asia: Thermal and

Mechanical Roles for the Tibetan Plateau. *Annual Review of Earth and Planetary Sciences*, *38*(1), 77–102. https://doi.org/10.1146/annurev-earth-040809-152456

Morlighem, M., Williams, C. N., Rignot, E., An, L., Arndt, J. E., Bamber, J. L., … Zinglersen, K. B. (2017). BedMachine v3: Complete Bed Topography and Ocean Bathymetry Mapping of Greenland From Multibeam Echo Sounding Combined With Mass Conservation. *Geophysical Research Letters*, *44*(21), 11,11-51,61. https://doi.org/https://doi.org/10.1002/2017GL074954

Muglia, J., Skinner, L. C., & Schmittner, A. (2018). Weak overturning circulation and high Southern Ocean nutrient utilization maximized glacial ocean carbon. *Earth and Planetary Science Letters*, *496*, 47–56. https://doi.org/https://doi.org/10.1016/j.epsl.2018.05.038

Murakami, T., Chen, L.-X., & Xie, A. (1986). Relationship among Seasonal Cycles, Low-Frequency Oscillations, and Transient Disturbances as Revealed from Outgoing Longwave Radiation Data. *Monthly Weather Review*, *114*(8), 1456–1465. https://doi.org/10.1175/1520-0493(1986)114<1456:RASCLF>2.0.CO;2

Mykleby, P. M., Snyder, P. K., & Twine, T. E. (2017). Quantifying the trade-off between carbon sequestration and albedo in midlatitude and high-latitude North American forests. *Geophysical Research Letters*, *44*(5), 2493–2501. https://doi.org/10.1002/2016GL071459

Nam, C., Bony, S., Dufresne, J. L., & Chepfer, H. (2012). The too few, too bright tropical low-cloud problem in CMIP5 models. *Geophysical Research Letters*, *39*(21), 1–7. https://doi.org/10.1029/2012GL053421

National Research Council. (2013). *Abrupt Impacts of Climate Change: Anticipating Surprises*. https://doi.org/10.17226/18373

Nepstad, D. C., Stickler, C. M., Soares-Filho, B., & Merry, F. (2008). Interactions among Amazon land use, forests and climate: Prospects for a near-term forest tipping point. *Philosophical Transactions of the Royal Society B: Biological Sciences*, *363*(1498), 1737–1746. https://doi.org/10.1098/rstb.2007.0036

Nepstad, D. C., Tohver, I. M., Ray, D., Moutinho, P., & Cardinot, G. (2007). MORTALITY OF LARGE TREES AND LIANAS FOLLOWING EXPERIMENTAL DROUGHT IN AN AMAZON FOREST. *Ecology*, *88*(9), 2259–2269. https://doi.org/10.1890/06-1046.1

Nepstad, D. C., Veríssimo, A., Alencar, A., Nobre, C., Lima, E., Lefebvre, P., … Brooks, V. (1999). Large-scale impoverishment of amazonian forests by logging and fire. *Nature*, *398*(6727), 505–508. https://doi.org/10.1038/19066

Nghiem, S. V., Rigor, I. G., Perovich, D. K., Clemente-Colón, P., Weatherly, J. W., & Neumann, G. (2007). Rapid reduction of Arctic perennial sea ice. *Geophysical Research Letters*, *34*(19), 1–6. https://doi.org/10.1029/2007GL031138

Nicholson, S. E., & Grist, J. P. (2001). A conceptual model for understanding rainfall variability in the West African Sahel on interannual and interdecadal timescales. *International Journal of Climatology*, *21*(14), 1733–1757. https://doi.org/https://doi.org/10.1002/joc.648

Nitzbon, J., Westermann, S., Langer, M., Martin, L. C. P., Strauss, J., Laboor, S., & Boike, J. (2020). Fast response of cold ice-rich permafrost in northeast Siberia to a warming climate. *Nature Communications*, *11*(1), 2201.

https://doi.org/10.1038/s41467-020-15725-8

Niyogi, D., Kishtawal, C., Tripathi, S., & Govindaraju, R. S. (2010). Observational evidence that agricultural intensification and land use change may be reducing the Indian summer monsoon rainfall. *Water Resources Research*, *46*(3). https://doi.org/https://doi.org/10.1029/2008WR007082

Nobre, Carlos A., Sampaio, G., Borma, L. S., Castilla-Rubio, J. C., Silva, J. S., & Cardoso, M. (2016). Land-use and climate change risks in the amazon and the need of a novel sustainable development paradigm. *Proceedings of the National Academy of Sciences of the United States of America*, *113*(39), 10759–10768. https://doi.org/10.1073/pnas.1605516113

Nobre, Carlos A., Sellers, P. J., & Shukla, J. (1991). Amazonian Deforestation and Regional Climate Change. *Journal of Climate*, *4*(10), 957–988. https://doi.org/10.1175/1520-0442(1991)004<0957:adarcc>2.0.co;2

Nobre, Carlos Afonso, & Borma, L. D. S. (2009). "Tipping points" for the Amazon forest. *Current Opinion in Environmental Sustainability*, *1*(1), 28–36. https://doi.org/10.1016/j.cosust.2009.07.003

Notz, D. (2009). The future of ice sheets and sea ice: Between reversible retreat and unstoppable loss. *Proceedings of the National Academy of Sciences of the United States of America*, *106*(49), 20590–20595. https://doi.org/10.1073/pnas.0902356106

Notz, D., & Marotzke, J. (2012). Observations reveal external driver for Arctic sea-ice retreat. *Geophysical Research Letters*, *39*(8). https://doi.org/10.1029/2012GL051094

Notz, D., & Stroeve, J. (2016). Observed Arctic sea-ice loss directly follows anthropogenic CO2 emission. *Science*, *354*(6313), 747–750. https://doi.org/10.1126/science.aag2345

Notz, D., & Stroeve, J. (2018). The Trajectory Towards a Seasonally Ice-Free Arctic Ocean. *Current Climate Change Reports*, *4*(4), 407–416. https://doi.org/10.1007/s40641-018-0113-2

Numaguti, A. (1995). Characteristics of 4-to-20-Day-Period Disturbances Observed in the Equatorial Pacific during the TOGA COARE IOP. *Journal of the Meteorological Society of Japan. Ser. II*, *73*(2B), 353–377. https://doi.org/10.2151/jmsj1965.73.2B_353

Olefeldt, D., Goswami, S., Grosse, G., Hayes, D., Hugelius, G., Kuhry, P., … Turetsky, M. R. (2016). Circumpolar distribution and carbon storage of thermokarst landscapes. *Nature Communications*, *7*, 1–11. https://doi.org/10.1038/ncomms13043

Oliveira, L. J. C., Costa, M. H., Soares-Filho, B. S., & Coe, M. T. (2013). Large-scale expansion of agriculture in Amazonia may be a no-win scenario. *Environmental Research Letters*, *8*(2). https://doi.org/10.1088/1748-9326/8/2/024021

Onarheim, I. H., Eldevik, T., Smedsrud, L. H., & Stroeve, J. C. (2018). Seasonal and regional manifestation of Arctic sea ice loss. *Journal of Climate*, *31*(12), 4917–4932. https://doi.org/10.1175/JCLI-D-17-0427.1

Oppenheimer, M., Glavovic, B., Hinkel, J., van de Wal, R., Magnan, A. K., Abd-Elgawad, A., … Sebesvari, Z. (2019). Sea Level Rise and Implications for Low Lying Islands, Coasts and Communities. In *IPCC Special Report on the Ocean and Cryosphere in a Changing Climate*. https://doi.org/10.1126/science.aam6284

Orr, J. C., Fabry, V. J., Aumont, O., Bopp, L., Doney, S. C., Feely, R. A., … Yool, A. (2005). Anthropogenic ocean acidification over the twenty-first century and its impact on calcifying organisms. *Nature*, *437*(7059), 681–686.

https://doi.org/10.1038/nature04095

Overland, J. E., Wang, M., Walsh, J. E., Christensen, J. H., Kattsov, V. M., & Chapman, W. L. (2011). Chapter 3: Climate Model Projections for the Arctic. *Snow, Water, Ice and Permafrost in the Arctic (SWIPA): Climate Change and the Cryosphere*, 37–54.

Palviainen, M., Laurén, A., Pumpanen, J., Bergeron, Y., Bond-Lamberty, B., Larjavaara, M., … Berninger, F. (2020). Decadal-Scale Recovery of Carbon Stocks After Wildfires Throughout the Boreal Forests. *Global Biogeochemical Cycles*, *34*(8), e2020GB006612. https://doi.org/https://doi.org/10.1029/2020GB006612

Pan, L., Powell, E. M., Latychev, K., Mitrovica, J. X., Creveling, J. R., Gomez, N., … Clark, P. U. (2021). Rapid postglacial rebound amplifies global sea level rise following West Antarctic Ice Sheet collapse. *Science Advances*, *7*(18), eabf7787. https://doi.org/10.1126/sciadv.abf7787

Pan, Y., Birdsey, R. A., Fang, J., Houghton, R., Kauppi, P. E., Kurz, W. A., … Hayes, D. (2011). A large and persistent carbon sink in the world's forests. *Science*, *333*(6045), 988–993. https://doi.org/10.1126/science.1201609

Paolo, F. S., Fricker, H. A., & Padman, L. (2015). Volume loss from Antarctic ice shelves is accelerating. *Science*, *348*(6232), 327–331. https://doi.org/10.1126/science.aaa0940

Parsons, L A. (2020). Implications of CMIP6 Projected Drying Trends for 21st Century Amazonian Drought Risk. *Earth's Future*, *8*(10), e2020EF001608. https://doi.org/https://doi.org/10.1029/2020EF001608

Parsons, Luke A., Yin, J., Overpeck, J. T., Stouffer, R. J., & Malyshev, S. (2014). Influence of the atlantic meridional overturning circulation on the monsoon rainfall and carbon balance of the American tropics. *Geophysical Research Letters*, *41*(1), 146–151. https://doi.org/10.1002/2013GL058454

Pascale, S., Carvalho, L. M. V, Adams, D. K., Castro, C. L., & Cavalcanti, I. F. A. (2019). Current and Future Variations of the Monsoons of the Americas in a Warming Climate. *Current Climate Change Reports*, *5*(3), 125–144. https://doi.org/10.1007/s40641-019-00135-w

Pastick, N. J., Jorgenson, M. T., Goetz, S. J., Jones, B. M., Wylie, B. K., Minsley, B. J., … Jorgenson, J. C. (2019). Spatiotemporal remote sensing of ecosystem change and causation across Alaska. *Global Change Biology*, *25*(3), 1171–1189. https://doi.org/10.1111/gcb.14279

Pattyn, F., Ritz, C., Hanna, E., Asay-Davis, X., DeConto, R., Durand, G., … van den Broeke, M. (2018, December 1). The Greenland and Antarctic ice sheets under 1.5 °C global warming. *Nature Climate Change*, Vol. 8, pp. 1053–1061. https://doi.org/10.1038/s41558-018-0305-8

Paul, S., Ghosh, S., Oglesby, R., Pathak, A., Chandrasekharan, A., & Ramsankaran, R. (2016). Weakening of Indian Summer Monsoon Rainfall due to Changes in Land Use Land Cover. *Scientific Reports*, *6*(1), 32177. https://doi.org/10.1038/srep32177

Pausata, F. S. R., Battisti, D. S., Nisancioglu, K. H., & Bitz, C. M. (2011). Chinese stalagmite δ18O controlled by changes in the Indian monsoon during a simulated Heinrich event. *Nature Geoscience*, *4*(7), 474–480. https://doi.org/10.1038/ngeo1169

Pearson, R. G., Phillips, S. J., Loranty, M. M., Beck, P. S. A., Damoulas, T., Knight, S. J., & Goetz, S. J. (2013). Shifts in Arctic vegetation and associated feedbacks under climate change. *Nature Climate Change*, *3*(7), 673–677. https://doi.org/10.1038/nclimate1858

Perovich, D. K., Light, B., Eicken, H., Jones, K. F., Runciman, K., & Nghiem, S. V. (2007). Increasing solar heating of the Arctic Ocean and adjacent seas, 1979-2005: Attribution and role in the ice-albedo feedback. *Geophysical Research Letters*, *34*(19), 1–5. https://doi.org/10.1029/2007GL031480

Perry, C. T., Alvarez-Filip, L., Graham, N. A. J., Mumby, P. J., Wilson, S. K., Kench, P. S., … MacDonald, C. (2018). Loss of coral reef growth capacity to track future increases in sea level. *Nature*, *558*(7710), 396–400. https://doi.org/10.1038/s41586-018-0194-z

Peters, G. P., Andrew, R. M., Canadell, J. G., Friedlingstein, P., Jackson, R. B., Korsbakken, J. I., … Peregon, A. (2019). Carbon dioxide emissions continue to grow amidst slowly emerging climate policies. *Nature Climate Change*. https://doi.org/10.1038/s41558-019-0659-6

Peters, G. P., Andrew, R. M., Canadell, J. G., Friedlingstein, P., Jackson, R. B., Korsbakken, J. I., … Peregon, A. (2020). Carbon dioxide emissions continue to grow amidst slowly emerging climate policies. *Nature Climate Change*, Vol. 10, pp. 3–6. https://doi.org/10.1038/s41558-019-0659-6

Petrenko, V. V., Smith, A. M., Schaefer, H., Riedel, K., Brook, E., Baggenstos, D., … Severinghaus, J. P. (2017). Minimal geological methane emissions during the Younger Dryas-Preboreal abrupt warming event. *Nature*, *548*(7668), 443–446. https://doi.org/10.1038/nature23316

Phillips, O. L., Aragão, L. E. O. C., Lewis, S. L., Fisher, J. B., Lloyd, J., López-González, G., … Torres-Lezama, A. (2009). Drought sensitivity of the amazon rainforest. *Science*, *323*(5919), 1344–1347. https://doi.org/10.1126/science.1164033

Phillips, O. L., van der Heijden, G., Lewis, S. L., López-González, G., Aragão, L. E. O. C., Lloyd, J., … Vilanova, E. (2010). Drought–mortality relationships for tropical forests. *New Phytologist*, *187*(3), 631–646. https://doi.org/https://doi.org/10.1111/j.1469-8137.2010.03359.x

Piao, S., Wang, X., Park, T., Chen, C., Lian, X., He, Y., … Myneni, R. B. (2020). Characteristics, drivers and feedbacks of global greening. *Nature Reviews Earth & Environment*, *1*(January). https://doi.org/10.1038/s43017-019-0001-x

Pickart, R. S., & Spall, M. A. (2007). Impact of Labrador Sea convection on the North Atlantic meridional overturning circulation. *Journal of Physical Oceanography*, *37*(9), 2207–2227. https://doi.org/10.1175/JPO3178.1

Piñero, E., Marquardt, M., Hensen, C., Haeckel, M., & Wallmann, K. (2013). Estimation of the global inventory of methane hydrates in marine sediments using transfer functions. *Biogeosciences*, *10*(2), 959–975. https://doi.org/10.5194/bg-10-959-2013

Pistone, K., Eisenman, I., & Ramanathan, V. (2019). Radiative Heating of an Ice-Free Arctic Ocean. *Geophysical Research Letters*, *46*(13), 7474–7480. https://doi.org/10.1029/2019GL082914

Plumb, R. A., & Hou, A. Y. (1992). The Response of a Zonally Symmetric Atmosphere to Subtropical Thermal Forcing: Threshold Behavior. *Journal of Atmospheric Sciences*, *49*(19), 1790–1799. https://doi.org/10.1175/1520-

0469(1992)049<1790:TROAZS>2.0.CO;2

Pollard, D., DeConto, R. M., & Alley, R. B. (2015). Potential Antarctic Ice Sheet retreat driven by hydrofracturing and ice cliff failure. *Earth and Planetary Science Letters*, *412*, 112–121. https://doi.org/10.1016/j.epsl.2014.12.035

Pueyo, S., de Alencastro Graça, P. M. L., Barbosa, R. I., Cots, R., Cardona, E., & Fearnside, P. M. (2010). Testing for criticality in ecosystem dynamics: The case of Amazonian rainforest and savanna fire. *Ecology Letters*, *13*(7), 793–802. https://doi.org/10.1111/j.1461-0248.2010.01497.x

Putnam, H. M. (2021). Avenues of reef-building coral acclimatization in response to rapid environmental change. *Journal of Experimental Biology*, *224*(Suppl_1). https://doi.org/10.1242/jeb.239319

Qin, Y., Xiao, X., Wigneron, J.-P., Ciais, P., Brandt, M., Fan, L., … Moore, B. (2021). Carbon loss from forest degradation exceeds that from deforestation in the Brazilian Amazon. *Nature Climate Change*, *11*(5), 442–448. https://doi.org/10.1038/s41558-021-01026-5

Raftery, A. E., Zimmer, A., Frierson, D. M. W., Startz, R., & Liu, P. (2017). Less than 2 °c warming by 2100 unlikely. *Nature Climate Change*, *7*(9), 637–641. https://doi.org/10.1038/nclimate3352

Rahmstorf, S., & England, M. H. (1997). On the influence of Southern Hemisphere winds on North Atlantic Deep Water flow. *PIK Report*, *26*(1988), 2040–2054.

Rahmstorf, S. (1996). On the freshwater forcing and transport of the Atlantic thermohaline circulation. *Climate Dynamics*, *12*(12), 799–811. https://doi.org/10.1007/s003820050144

Rahmstorf, Stefan. (2002, September 12). Ocean circulation and climate during the past 120,000 years. *Nature*, Vol. 419, pp. 207–214. https://doi.org/10.1038/nature01090

Rahmstorf, Stefan, Box, J. E., Feulner, G., Mann, M. E., Robinson, A., Rutherford, S., & Schaffernicht, E. J. (2015). Exceptional twentieth-century slowdown in Atlantic Ocean overturning circulation. *Nature Climate Change*, *5*(5), 475–480. https://doi.org/10.1038/nclimate2554

Ramanathan, V., Chung, C., Kim, D., Bettge, T., Buja, L., Kiehl, J. T., … Wild, M. (2005). Atmospheric brown clouds: Impacts on South Asian climate and hydrological cycle. *Proceedings of the National Academy of Sciences of the United States of America*, *102*(15), 5326 LP – 5333. https://doi.org/10.1073/pnas.0500656102

Ramaswamy, V., Boucher, O., Haigh, J., Hauglustaine, D., Haywood, J., Myhre, G., … Solomon, S. (2001). Radiative forcing of climate change. Climate change 2001: the scientific basis. Contribution of working group I to the third assessment report of the intergovernmental panel on climate change. *DJ Griggs, M Noguer, PJ van Der Linden, X Dai, K Maskell and CA Johnson (Cambridge: Cambridge University Press) Pp*, *350*, 416.

Randerson, J. T., Liu, H., Flanner, M. G., Chambers, S. D., Jin, Y., Hess, P. G., … Zender, C. S. (2006). The impact of boreal forest fire on climate warming. *Science*, *314*(5802), 1130–1132. https://doi.org/10.1126/science.1132075

Rasmussen, D. J., Buchanan, M. K., Kopp, R. E., & Oppenheimer, M. (2020). A Flood Damage Allowance Framework for Coastal Protection With Deep Uncertainty in Sea Level Rise. *Earth's Future*, *8*(3), 1–16. https://doi.org/10.1029/2019EF001340

Reeburgh, W. S. (2007). *Oceanic Methane Biogeochemistry*. https://doi.org/10.1021/cr050362v

Rees, W. G., Hofgaard, A., Boudreau, S., Cairns, D. M., Harper, K., Mamet, S., … Tutubalina, O. (2020). Is subarctic forest advance able to keep pace with climate change? *Global Change Biology*, *26*(7), 3965–3977. https://doi.org/https://doi.org/10.1111/gcb.15113

Riahi, K., van Vuuren, D. P., Kriegler, E., Edmonds, J., O'Neill, B. C., Fujimori, S., … Tavoni, M. (2017). The Shared Socioeconomic Pathways and their energy, land use, and greenhouse gas emissions implications: An overview. *Global Environmental Change*, *42*, 153–168. https://doi.org/10.1016/j.gloenvcha.2016.05.009

Ridley, J. K., Lowe, J. A., & Hewitt, H. T. (2012). How reversible is sea ice loss? *Cryosphere*, *6*(1), 193–198. https://doi.org/10.5194/tc-6-193-2012

Rignot, E., Mouginot, J., Morlighem, M., Seroussi, H., & Scheuchl, B. (2014). Widespread, rapid grounding line retreat of Pine Island, Thwaites, Smith, and Kohler glaciers, West Antarctica, from 1992 to 2011. *Geophysical Research Letters*, *41*(10), 3502–3509. https://doi.org/10.1002/2014GL060140

Ritchie, P. D. L., Clarke, J. J., Cox, P. M., & Huntingford, C. (2021). Overshooting tipping point thresholds in a changing climate. *Nature*, *592*(7855), 517–523. https://doi.org/10.1038/s41586-021-03263-2

Ritson-Williams, R., & Gates, R. D. (2020). Coral community resilience to successive years of bleaching in Kāneʻohe Bay, Hawaiʻi. *Coral Reefs*, *39*(3), 757–769. https://doi.org/10.1007/s00338-020-01944-4

Rivkina, E., Laurinavichius, K., McGrath, J., Tiedje, J., Shcherbakova, V., & Gilichinsky, D. (2004). Microbial life in permafrost. *Advances in Space Research*, *33*(8), 1215–1221. https://doi.org/10.1016/j.asr.2003.06.024

Roberts, C. D., Waters, J., Peterson, K. A., Palmer, M. D., McCarthy, G. D., Frajka-Williams, E., … Zuo, H. (2013). Atmosphere drives recent interannual variability of the Atlantic meridional overturning circulation at 26.5°N. *Geophysical Research Letters*, *40*(19), 5164–5170. https://doi.org/https://doi.org/10.1002/grl.50930

Roberts, M. J., Jackson, L. C., Roberts, C. D., Meccia, V., Docquier, D., Koenigk, T., … Wu, L. (2020). Sensitivity of the Atlantic Meridional Overturning Circulation to Model Resolution in CMIP6 HighResMIP Simulations and Implications for Future Changes. *Journal of Advances in Modeling Earth Systems*, *12*(8), e2019MS002014. https://doi.org/https://doi.org/10.1029/2019MS002014

Robinson, A., Calov, R., & Ganopolski, A. (2012). Multistability and critical thresholds of the Greenland ice sheet. *Nature Climate Change*, *2*(6), 429–432. https://doi.org/10.1038/nclimate1449

Rogers, B. M., Soja, A. J., Goulden, M. L., & Randerson, J. T. (2015). Influence of tree species on continental differences in boreal fires and climate feedbacks. *Nature Geoscience*, *8*(3), 228–234. https://doi.org/10.1038/ngeo2352

Rolph, R. J., Feltham, D. L., & Schröder, D. (2020). Changes of the Arctic marginal ice zone during the satellite era. *The Cryosphere*, *14*(6), 1971–1984. https://doi.org/10.5194/tc-14-1971-2020

Romanovsky, V. E., Smith, S. L., Shiklomanov, N. I., Streletskiy, D. A., Isaksen, K., Kholodov, A. L., … Marchenko, S. S. (2017). Terrestrial permafrost. In J. Blunden & D. Arndt (Eds.), *State of the Climate in 2016*. Boston, MA: Bulletin of the American Meterological Society.

Romanovsky, V., Isaksen, K., D, D., Anisimov, O., A, I., Leibman, M., … D, W. (2017). *Changing permafrost and its impacts. In: Snow, Water, Ice and Permafrost in the Arctic (SWIPA) 2017*.

Ross, N., Bingham, R. G., Corr, H. F. J., Ferraccioli, F., Jordan, T. A., Le Brocq, A., … Siegert, M. J. (2012). Steep reverse bed slope at the grounding line of the Weddell Sea sector in West Antarctica. *Nature Geoscience*, *5*(6), 393–396. https://doi.org/10.1038/ngeo1468

Rotstayn, L. D., & Lohmann, U. (2002). Tropical rainfall trends and the indirect aerosol effect. *Journal of Climate*, *15*(15), 2103–2116. https://doi.org/10.1175/1520-0442(2002)015<2103:TRTATI>2.0.CO;2

Roxy, M. K., Ritika, K., Terray, P., Murtugudde, R., Ashok, K., & Goswami, B. N. (2015). Drying of Indian subcontinent by rapid Indian Ocean warming and a weakening land-sea thermal gradient. *Nature Communications*, *6*(1), 7423. https://doi.org/10.1038/ncomms8423

Rudy, A. C. A., Lamoureux, S. F., Kokelj, S. V., Smith, I. R., & England, J. H. (2017). Accelerating Thermokarst Transforms Ice-Cored Terrain Triggering a Downstream Cascade to the Ocean. *Geophysical Research Letters*, *44*(21), 11,080-11,087. https://doi.org/10.1002/2017GL074912

Ruppel, C D, & Waite, W. F. (2020). Timescales and Processes of Methane Hydrate Formation and Breakdown, With Application to Geologic Systems. *Journal of Geophysical Research: Solid Earth*, *125*(8), e2018JB016459. https://doi.org/https://doi.org/10.1029/2018JB016459

Ruppel, Carolyn D., & Kessler, J. D. (2017, March 1). The interaction of climate change and methane hydrates. *Reviews of Geophysics*, Vol. 55, pp. 126–168. https://doi.org/10.1002/2016RG000534

Ruppel, Carolyn D. (2011). Methane hydrates and contemporary climate change. *Nature Eduction Knowledge*, *2*(12), 12-. Retrieved from http://pubs.er.usgs.gov/publication/70007207

Ryan, J. C., Smith, L. C., Van As, D., Cooley, S. W., Cooper, M. G., Pitcher, L. H., & Hubbard, A. (2019). Greenland Ice Sheet surface melt amplified by snowline migration and bare ice exposure. *Science Advances*, *5*(3). https://doi.org/10.1126/sciadv.aav3738

Saatchi, S., Asefi-Najafabady, S., Malhi, Y., Aragão, L. E. O. C., Anderson, L. O., Myneni, R. B., & Nemani, R. (2013). Persistent effects of a severe drought on Amazonian forest canopy. *Proceedings of the National Academy of Sciences*, *110*(2), 565–570. https://doi.org/10.1073/pnas.1204651110

Saatchi, S. S., Harris, N. L., Brown, S., Lefsky, M., Mitchard, E. T. A., Salas, W., … Morel, A. (2011). Benchmark map of forest carbon stocks in tropical regions across three continents. *Proceedings of the National Academy of Sciences of the United States of America*, *108*(24), 9899–9904. https://doi.org/10.1073/pnas.1019576108

Saha, S., Chakraborty, D., Paul, R. K., Samanta, S., & Singh, S. B. (2018). Disparity in rainfall trend and patterns among different regions: analysis of 158 years' time series of rainfall dataset across India. *Theoretical and Applied Climatology*, *134*(1), 381–395. https://doi.org/10.1007/s00704-017-2280-9

Salati, E., Dall'Olio, A., Matsui, E., & Gat, J. R. (1979). Recycling of water in the Amazon Basin: An isotopic study. *Water Resources Research*, *15*(5), 1250–1258. https://doi.org/10.1029/WR015i005p01250

Salati, E., & Vose, P. B. (1984). Amazon Basin: A system in equilibrium. *Science*, *225*(4658), 129–138. https://doi.org/10.1126/science.225.4658.129

Salazar, L. F., & Nobre, C. A. (2010). Climate change and thresholds of biome shifts in Amazonia. *Geophysical Research Letters*, *37*(17), n/a-n/a. https://doi.org/10.1029/2010GL043538

Sampe, T., & Xie, S.-P. (2010). Large-Scale Dynamics of the Meiyu-Baiu Rainband: Environmental Forcing by the Westerly Jet. *Journal of Climate*, *23*(1), 113–134. https://doi.org/10.1175/2009JCLI3128.1

Sassen, R., Joye, S., Sweet, S. T., Defreitas, D. A., Milkov, A. V., & MacDonald, I. R. (1999). Thermogenic gas hydrates and hydrocarbon gases in complex chemosynthetic communities, Gulf of Mexico continental slope. *Organic Geochemistry*, *30*(7), 485–497. https://doi.org/10.1016/S0146-6380(99)00050-9

Saunois, M, Stavert, A. R., Poulter, B., Bousquet, P., Canadell, J. G., Jackson, R. B., … Zhuang, Q. (2020). The Global Methane Budget 2000--2017. *Earth System Science Data*, *12*(3), 1561–1623. https://doi.org/10.5194/essd-12-1561-2020

Saunois, Marielle, Bousquet, P., Poulter, B., Peregon, A., Ciais, P., Canadell, J. G., … Zhu, Q. (2016). The global methane budget 2000-2012. *Earth System Science Data*, *8*(2), 697–751. https://doi.org/10.5194/essd-8-697-2016

Sayedi, S. S., Abbott, B. W., Thornton, B. F., Frederick, J. M., Vonk, J. E., Overduin, P., … Frei, R. J. (2020). Subsea permafrost carbon stocks and climate change sensitivity estimated by expert assessment. *Environmental Research Letters*, *15*(12), 124075. https://doi.org/10.1088/1748-9326/abcc29

Schädel, C., Bader, M. K.-F., Schuur, E. A. G., Biasi, C., Bracho, R., Čapek, P., … Wickland, K. P. (2016). Potential carbon emissions dominated by carbon dioxide from thawed permafrost soils. *Nature Climate Change*, *6*(10), 950–953. https://doi.org/10.1038/nclimate3054

Schaefer, K., Zhang, T., Bruhwiler, L., & Barrett, A. P. (2011). Amount and timing of permafrost carbon release in response to climate warming. *Tellus B: Chemical and Physical Meteorology*, *63*(2), 168–180. https://doi.org/10.1111/j.1600-0889.2010.00527.x

Schaphoff, S., Heyder, U., Ostberg, S., Gerten, D., Heinke, J., & Lucht, W. (2013). Contribution of permafrost soils to the global carbon budget. *Environmental Research Letters*, *8*(1). https://doi.org/10.1088/1748-9326/8/1/014026

Schaphoff, S., Reyer, C. P. O., Schepaschenko, D., Gerten, D., & Shvidenko, A. (2016, February 1). Tamm Review: Observed and projected climate change impacts on Russia's forests and its carbon balance. *Forest Ecology and Management*, Vol. 361, pp. 432–444. https://doi.org/10.1016/j.foreco.2015.11.043

Scheffer, M., Hirota, M., Holmgren, M., Van Nes, E. H., & Chapin, F. S. (2012). Thresholds for boreal biome transitions. *Proceedings of the National Academy of Sciences of the United States of America*, *109*(52), 21384–21389. https://doi.org/10.1073/pnas.1219844110

Schewe, J., Levermann, A., & Cheng, H. (2012). A critical humidity threshold for monsoon transitions. *Climate of the Past*, *8*(2), 535–544. https://doi.org/10.5194/cp-8-535-2012

Schleussner, C F, Frieler, K., Meinshausen, M., Yin, J., & Levermann, A. (2011). Emulating Atlantic overturning strength for low emission scenarios: Consequences for sea-level rise along the North American east coast. *Earth System Dynamics*, *2*(2), 191–200. https://doi.org/10.5194/esd-2-191-2011

Schleussner, Carl Friedrich, Lissner, T. K., Fischer, E. M., Wohland, J., Perrette, M., Golly, A., … Schaeffer, M. (2016). Differential climate impacts for policy-relevant limits to global warming: The case of 1.5 °c and 2 °c. *Earth System Dynamics*, *7*(2), 327–351. https://doi.org/10.5194/esd-7-327-2016

Schneider, T., Kaul, C. M., & Pressel, K. G. (2019). Possible climate transitions from breakup of stratocumulus decks under greenhouse warming. *Nature Geoscience*, *12*(3), 164–168. https://doi.org/10.1038/s41561-019-0310-1

Schneider, T., Kaul, C. M., & Pressel, K. G. (2020). Solar geoengineering may not prevent strong warming from direct effects of CO₂ on stratocumulus cloud cover. *Proceedings of the National Academy of Sciences*, *117*(48), 30179 LP – 30185. https://doi.org/10.1073/pnas.2003730117

Schneider Von Deimling, T., Grosse, G., Strauss, J., Schirrmeister, L., Morgenstern, A., Schaphoff, S., … Boike, J. (2015). Observation-based modelling of permafrost carbon fluxes with accounting for deep carbon deposits and thermokarst activity. *Biogeosciences*, *12*(11), 3469–3488. https://doi.org/10.5194/bg-12-3469-2015

Schneider von Deimling, T., Meinshausen, M., Levermann, A., Huber, V., Frieler, K., Lawrence, D. M., & Brovkin, V. (2012). Estimating the near-surface permafrost-carbon feedback on global warming. *Biogeosciences*, *9*(2), 649–665. https://doi.org/10.5194/bg-9-649-2012

Schoof, C. (2007). Ice sheet grounding line dynamics: Steady states, stability, and hysteresis. *Journal of Geophysical Research*, *112*(F3), F03S28. https://doi.org/10.1029/2006JF000664

Schuur, E. A. G., McGuire, A. D., Romanovsky, C., Schädel, C., & Mack, M. (2018). *Chapter 11: Arctic and boreal carbon. In Second State of the Carbon Cycle Report (SOCCR2): A Sustained Assessment Report [Cavallaro, N., G. Shrestha, R. Birdsey, M. A. Mayes, R. G. Najjar, S. C. Reed, P. Romero-Lankao, and Z. Zhu (eds.)]*. https://doi.org/https://doi.org/10.7930/ SOCCR2.2018.Ch11

Schuur, E. A.G., McGuire, A. D., Schädel, C., Grosse, G., Harden, J. W., Hayes, D. J., … Vonk, J. E. (2015, April 9). Climate change and the permafrost carbon feedback. *Nature*, Vol. 520, pp. 171–179. https://doi.org/10.1038/nature14338

Schuur, Edward A.G., & Abbott, B. (2011, December 1). Climate change: High risk of permafrost thaw. *Nature*, Vol. 480, pp. 32–33. https://doi.org/10.1038/480032a

Schwietzke, S., Sherwood, O. A., Bruhwiler, L. M. P., Miller, J. B., Etiope, G., Dlugokencky, E. J., … Tans, P. P. (2016). Upward revision of global fossil fuel methane emissions based on isotope database. *Nature*, *538*(7623), 88–91. https://doi.org/10.1038/nature19797

Screen, J. A., Deser, C., Simmonds, I., & Tomas, R. (2014). Atmospheric impacts of Arctic sea-ice loss, 1979-2009: Separating forced change from atmospheric internal variability. *Climate Dynamics*, *43*(1–2), 333–344. https://doi.org/10.1007/s00382-013-1830-9

Screen, J. A., & Simmonds, I. (2010). The central role of diminishing sea ice in recent Arctic temperature amplification. *Nature*, *464*(7293), 1334–1337. https://doi.org/10.1038/nature09051

Seidl, R., Honkaniemi, J., Aakala, T., Aleinikov, A., Angelstam, P., Bouchard, M., … Senf, C. (2020). Globally consistent climate sensitivity of natural disturbances across boreal and temperate forest ecosystems. *Ecography*, *43*(7), 967–978. https://doi.org/https://doi.org/10.1111/ecog.04995

Seneca Lindsey, D., & Dupont, T. K. (2012). Mechanical effect of mélange-induced buttressing on embayment-terminating glacier dynamics. *The Cryosphere Discussions*, *6*(5), 4123–4136. https://doi.org/10.5194/tcd-6-4123-2012

Seroussi, H., Nowicki, S., Payne, A., Goelzer, H., Lipscomb, W., Abe Ouchi, A., … Zwinger, T. (2020). ISMIP6 Antarctica: a multi-model ensemble of the Antarctic ice sheet evolution over the 21st century. *The Cryosphere Discussions*, 1–54. https://doi.org/10.5194/tc-2019-324

Serreze, M. C., Barrett, A. P., Stroeve, J. C., Kindig, D. N., & Holland, M. M. (2009). The emergence of surface-based Arctic amplification. *Cryosphere*, *3*(1), 11–19. https://doi.org/10.5194/tc-3-11-2009

Seshadri, A. K. (2017). Energetics and monsoon bifurcations. *Climate Dynamics*, *48*(1), 561–576. https://doi.org/10.1007/s00382-016-3094-7

Shakhova, N., Semiletov, I., Leifer, I., Sergienko, V., Salyuk, A., Kosmach, D., … Gustafsson, Ö. (2014). Ebullition and storm-induced methane release from the East Siberian Arctic Shelf. *Nature Geoscience*, *7*(1), 64–70. https://doi.org/10.1038/ngeo2007

Shanahan, T. M., McKay, N. P., Hughen, K. A., Overpeck, J. T., Otto-Bliesner, B., Heil, C. W., … Peck, J. (2015). The time-transgressive termination of the African Humid Period. *Nature Geoscience*, *8*(2), 140–144. https://doi.org/10.1038/ngeo2329

Shen, Z., Duan, A., Li, D., & Li, J. (2021). Assessment and Ranking of Climate Models in Arctic Sea Ice Cover Simulation: From CMIP5 to CMIP6. *Journal of Climate*, *34*(9), 3609–3627. https://doi.org/10.1175/JCLI-D-20-0294.1

Shepherd, A., Wingham, D., & Rignot, E. (2004). Warm ocean is eroding West Antarctic Ice Sheet. *Geophysical Research Letters*, *31*(23). https://doi.org/10.1029/2004GL021106

Sherriff, R. L., Berg, E. E., & Miller, A. E. (2011). Climate variability and spruce beetle (Dendroctonus rufipennis) outbreaks in south-central and southwest Alaska. *Ecology*, *92*(7), 1459–1470. https://doi.org/https://doi.org/10.1890/10-1118.1

Sherwood, S. C., Webb, M. J., Annan, J. D., Armour, K. C., Forster, P. M., Hargreaves, J. C., … Zelinka, M. D. (2020). An Assessment of Earth’s Climate Sensitivity Using Multiple Lines of Evidence. *Reviews of Geophysics*, *58*(4), e2019RG000678. https://doi.org/https://doi.org/10.1029/2019RG000678

Shugart, H. H., Asner, G. P., Fischer, R., Huth, A., Knapp, N., Le Toan, T., & Shuman, J. K. (2015). Computer and remote-sensing infrastructure to enhance large-scale testing of individual-based forest models. *Frontiers in Ecology and the Environment*, *13*(9), 503–511. https://doi.org/https://doi.org/10.1890/140327

Shukla, S. P., Puma, M. J., & Cook, B. I. (2014). The response of the South Asian Summer Monsoon circulation to intensified irrigation in global climate model simulations. *Climate Dynamics*, *42*(1), 21–36. https://doi.org/10.1007/s00382-013-1786-9

Shuman, J. K., Foster, A. C., Shugart, H. H., Hoffman-Hall, A., Krylov, A., Loboda, T., … Sochilova, E. (2017). Fire disturbance and climate change: Implications for Russian forests. *Environmental Research Letters*, *12*(3).

https://doi.org/10.1088/1748-9326/aa5eed

Shuman, J. K., Tchebakova, N. M., Parfenova, E. I., Soja, A. J., Shugart, H. H., Ershov, D., & Holcomb, K. (2015). Forest forecasting with vegetation models across Russia. *Canadian Journal of Forest Research*, *45*(2), 175–184. https://doi.org/10.1139/cjfr-2014-0138

Shur, Y. L., & Jorgenson, M. T. (2007). Patterns of permafrost formation and degradation in relation to climate and ecosystems. *Permafrost and Periglacial Processes*, *18*(1), 7–19. https://doi.org/10.1002/ppp.582

Shvidenko, A. Z., Shchepashchenko, D. G., Vaganov, E. A., Sukhinin, A. I., Maksyutov, S. S., McCallum, I., & Lakyda, I. P. (2011). Impact of wildfire in Russia between 1998-2010 on ecosystems and the global carbon budget. *Doklady Earth Sciences*, *441*(2), 1678–1682. https://doi.org/10.1134/S1028334X11120075

Silbiger, N. J., Guadayol, Ò., Thomas, F. I. M., & Donahue, M. J. (2014). Reefs shift from net accretion to net erosion along a natural environmental gradient. *Marine Ecology Progress Series*, *515*, 33–44. https://doi.org/10.3354/meps10999

Silva, C. V. J., Aragão, L. E. O. C., Young, P. J., Espirito-Santo, F., Berenguer, E., Anderson, L. O., … Barlow, J. (2020). Estimating the multi-decadal carbon deficit of burned Amazonian forests. *Environmental Research Letters*, *15*(11), 114023. https://doi.org/10.1088/1748-9326/abb62c

Silva Junior, C. H. L., Anderson, L. O., Silva, A. L., Almeida, C. T., Dalagnol, R., Pletsch, M. A. J. S., … Aragão, L. E. O. C. (2019). Fire Responses to the 2010 and 2015/2016 Amazonian Droughts. *Frontiers in Earth Science*, *7*, 97. https://doi.org/10.3389/feart.2019.00097

Silva Junior, C. H. L., Aragão, L. E. O. C., Anderson, L. O., Fonseca, M. G., Shimabukuro, Y. E., Vancutsem, C., … Saatchi, S. S. (2020). Persistent collapse of biomass in Amazonian forest edges following deforestation leads to unaccounted carbon losses. *Science Advances*, *6*(40), eaaz8360. https://doi.org/10.1126/sciadv.aaz8360

Silva Junior, C. H. L., Pessôa, A. C. M., Carvalho, N. S., Reis, J. B. C., Anderson, L. O., & Aragão, L. E. O. C. (2021). The Brazilian Amazon deforestation rate in 2020 is the greatest of the decade. *Nature Ecology & Evolution*, *5*(2), 144–145. https://doi.org/10.1038/s41559-020-01368-x

Silvano, A., Rintoul, S. R., & Herraiz-Borreguero, L. (2016). Ocean-ice shelf interaction in East Antarctica. *Oceanography*, *29*(4), 130–143. https://doi.org/10.5670/oceanog.2016.105

SIMIP Community. (2020). Arctic Sea Ice in CMIP6. *Geophysical Research Letters*, *47*(10), e2019GL086749. https://doi.org/https://doi.org/10.1029/2019GL086749

Singh, D., Ghosh, S., Roxy, M. K., & McDermid, S. (2019). Indian summer monsoon: Extreme events, historical changes, and role of anthropogenic forcings. *Wiley Interdisciplinary Reviews: Climate Change*, *10*(2), 1–35. https://doi.org/10.1002/wcc.571

Singh, D., Tsiang, M., Rajaratnam, B., & Diffenbaugh, N. S. (2014). Observed changes in extreme wet and dry spells during the south Asian summer monsoon season. *Nature Climate Change*, *4*(6), 456–461. https://doi.org/10.1038/nclimate2208

Slater, D. A., Felikson, D., Straneo, F., Goelzer, H., Little, C. M., Morlighem, M., … Nowicki, S. (2020). Twenty-first century ocean forcing of the Greenland ice sheet for modelling of sea level contribution. *The Cryosphere*, *14*(3), 985–1008.

https://doi.org/10.5194/tc-14-985-2020

Slater, T., Hogg, A. E., & Mottram, R. (2020). Ice-sheet losses track high-end sea-level rise projections. *Nature Climate Change*, *10*(10), 879–881. https://doi.org/10.1038/s41558-020-0893-y

Smith, C J, Forster, P. M., Allen, M., Leach, N., Millar, R. J., Passerello, G. A., & Regayre, L. A. (2018). FAIR v1.3: a simple emissions-based impulse response and carbon cycle model. *Geoscientific Model Development*, *11*(6), 2273–2297. https://doi.org/10.5194/gmd-11-2273-2018

Smith, C J, Kramer, R. J., Myhre, G., Alterskjær, K., Collins, W., Sima, A., … Forster, P. M. (2020). Effective radiative forcing and adjustments in CMIP6 models. *Atmospheric Chemistry and Physics*, *20*(16), 9591–9618. https://doi.org/10.5194/acp-20-9591-2020

Smith, Christopher J., Forster, P. M., Allen, M., Leach, N., Millar, R. J., Passerello, G. A., & Regayre, L. A. (2017). FAIR v1.1: A simple emissions-based impulse response and carbon cycle model. *Geoscientific Model Development Discussions*, (December), 1–45. https://doi.org/10.5194/gmd-2017-266

Soja, A. J., Tchebakova, N. M., French, N. H. F., Flannigan, M. D., Shugart, H. H., Stocks, B. J., … Stackhouse, P. W. (2007). Climate-induced boreal forest change: Predictions versus current observations. *Global and Planetary Change*, *56*(3–4), 274–296. https://doi.org/10.1016/j.gloplacha.2006.07.028

Solomon, S., Plattner, G. K., Knutti, R., & Friedlingstein, P. (2009). Irreversible climate change due to carbon dioxide emissions. *Proceedings of the National Academy of Sciences of the United States of America*, *106*(6), 1704–1709. https://doi.org/10.1073/pnas.0812721106

Spalding, M., Burke, L., Wood, S. A., Ashpole, J., Hutchison, J., & zu Ermgassen, P. (2017). Mapping the global value and distribution of coral reef tourism. *Marine Policy*, *82*(January), 104–113. https://doi.org/10.1016/j.marpol.2017.05.014

Spalding, M. D., & Brown, B. E. (2015, November 13). Warm-water coral reefs and climate change. *Science*, Vol. 350, pp. 769–771. https://doi.org/10.1126/science.aad0349

Spracklen, D. V, Arnold, S. R., & Taylor, C. M. (2012). Observations of increased tropical rainfall preceded by air passage over forests. *Nature*, *489*(7415), 282–285. https://doi.org/10.1038/nature11390

Srokosz, M. A., & Bryden, H. L. (2015, June 19). Observing the Atlantic Meridional Overturning Circulation yields a decade of inevitable surprises. *Science*, Vol. 348. https://doi.org/10.1126/science.1255575

Staal, A., Fetzer, I., Wang-Erlandsson, L., Bosmans, J. H. C., Dekker, S. C., van Nes, E. H., … Tuinenburg, O. A. (2020). Hysteresis of tropical forests in the 21st century. *Nature Communications*, *11*(1), 4978. https://doi.org/10.1038/s41467-020-18728-7

Staal, A., Flores, B. M., Aguiar, A. P. D., Bosmans, J. H. C., Fetzer, I., & Tuinenburg, O. A. (2020). Feedback between drought and deforestation in the Amazon. *Environmental Research Letters*, *15*(4). https://doi.org/10.1088/1748-9326/ab738e

Steffen, W., Rockström, J., Richardson, K., Lenton, T. M., Folke, C., Liverman, D., … Schellnhuber, H. J. (2018, August 14). Trajectories of the Earth System in the Anthropocene. *Proceedings of the National Academy of Sciences of the United States of America*, Vol. 115, pp. 8252–8259. https://doi.org/10.1073/pnas.1810141115

Steneck, R. S., Arnold, S. N., Boenish, R., de León, R., Mumby, P. J., Rasher, D. B., & Wilson, M. W. (2019). Managing Recovery Resilience in Coral Reefs Against Climate-Induced Bleaching and Hurricanes: A 15 Year Case Study From Bonaire, Dutch Caribbean. *Frontiers in Marine Science*, *6*, 265. https://doi.org/10.3389/fmars.2019.00265

Stickler, C. M., Coe, M. T., Costa, M. H., Nepstad, D. C., McGrath, D. G., Dias, L. C. P., … Soares-Filho, B. S. (2013). Dependence of hydropower energy generation on forests in the Amazon Basin at local and regional scales. *Proceedings of the National Academy of Sciences of the United States of America*, *110*(23), 9601–9606. https://doi.org/10.1073/pnas.1215331110

Stommel, H. (1961). Thermohaline Convection with Two Stable Regimes of Flow. *Tellus*, *13*(2), 224–230. https://doi.org/https://doi.org/10.1111/j.2153-3490.1961.tb00079.x

Storlazzi, C. D., Reguero, B. G., Cole, A. D., Lowe, E., Shope, J. B., Gibbs, A. E., … Beck, M. W. (2019). *Rigorously Valuing the Role of U . S . Coral Reefs in Coastal Hazard Risk Reduction: U.S. Geological Survey Open-File Report 2019–1027*. 42. https://doi.org/https://doi.org/10.3133/ofr20191027

Stralberg, D., Arseneault, D., Baltzer, J. L., Barber, Q. E., Bayne, E. M., Boulanger, Y., … Whitman, E. (2020). Climate-change refugia in boreal North America: what, where, and for how long? *Frontiers in Ecology and the Environment*, *18*(5), 261–270. https://doi.org/https://doi.org/10.1002/fee.2188

Stroeve, J., Barrett, A., Serreze, M., & Schweiger, A. (2014). Using records from submarine, aircraft and satellites to evaluate climate model simulations of Arctic sea ice thickness. *Cryosphere*, *8*(5), 1839–1854. https://doi.org/10.5194/tc-8-1839-2014

Stroeve, J. C., Kattsov, V., Barrett, A., Serreze, M., Pavlova, T., Holland, M., & Meier, W. N. (2012). Trends in Arctic sea ice extent from CMIP5, CMIP3 and observations. *Geophysical Research Letters*, *39*(16), 1–7. https://doi.org/10.1029/2012GL052676

Stroeve, J. C., Serreze, M. C., Holland, M. M., Kay, J. E., Malanik, J., & Barrett, A. P. (2012). The Arctic's rapidly shrinking sea ice cover: A research synthesis. *Climatic Change*, *110*(3–4), 1005–1027. https://doi.org/10.1007/s10584-011-0101-1

Stroeve, Julienne, Holland, M. M., Meier, W., Scambos, T., & Serreze, M. (2007). Arctic sea ice decline: Faster than forecast. *Geophysical Research Letters*, *34*(9), 1–5. https://doi.org/10.1029/2007GL029703

Stroeve, Julienne, & Notz, D. (2018, September 24). Changing state of Arctic sea ice across all seasons. *Environmental Research Letters*, Vol. 13. https://doi.org/10.1088/1748-9326/aade56

Sturm, M., & Massom, R. A. (2016). Snow in the sea ice system: Friend or foe? In *Sea Ice: Third Edition*. https://doi.org/10.1002/9781118778371.ch3

Sully, S., Burkepile, D. E., Donovan, M. K., Hodgson, G., & van Woesik, R. (2019). A global analysis of coral bleaching over the past two decades. *Nature Communications*, *10*(1). https://doi.org/10.1038/s41467-019-09238-2

Sultan, B., & Janicot, S. (2003). The West African Monsoon Dynamics. Part II: The ?Preonset? and ?Onset? of the Summer Monsoon. *Journal of Climate*, *16*(21), 3407–3427. https://doi.org/10.1175/1520-

0442(2003)016<3407:TWAMDP>2.0.CO;2

Sun, S., Pattyn, F., Simon, E. G., Albrecht, T., Cornford, S., Calov, R., … Zhang, T. (2020). Antarctic ice sheet response to sudden and sustained ice-shelf collapse (ABUMIP). *Journal of Glaciology*, *66*(260), 891–904. https://doi.org/DOI: 10.1017/jog.2020.67

Takahashi, T., Sutherland, S. C., Wanninkhof, R., Sweeney, C., Feely, R. A., Chipman, D. W., … de Baar, H. J. W. (2009). Climatological mean and decadal change in surface ocean pCO2, and net sea–air CO2 flux over the global oceans. *Deep Sea Research Part II: Topical Studies in Oceanography*, *56*(8), 554–577. https://doi.org/https://doi.org/10.1016/j.dsr2.2008.12.009

Takemura, T., Nozawa, T., Emori, S., Nakajima, T. Y., & Nakajima, T. (2005). Simulation of climate response to aerosol direct and indirect effects with aerosol transport-radiation model. *Journal of Geophysical Research: Atmospheres*, *110*(D2). https://doi.org/https://doi.org/10.1029/2004JD005029

Tan, Z., Schneider, T., Teixeira, J., & Pressel, K. G. (2017). Large-eddy simulation of subtropical cloud-topped boundary layers: 2. Cloud response to climate change. *Journal of Advances in Modeling Earth Systems*, *9*(1), 19–38. https://doi.org/10.1002/2016MS000804

Tape, K., Sturm, M., & Racine, C. (2006). The evidence for shrub expansion in Northern Alaska and the Pan-Arctic. *Global Change Biology*, *12*(4), 686–702. https://doi.org/10.1111/j.1365-2486.2006.01128.x

Taylor, C. M., Belušić, D., Guichard, F., Parker, D. J., Vischel, T., Bock, O., … Panthou, G. (2017). Frequency of extreme Sahelian storms tripled since 1982 in satellite observations. *Nature*, *544*(7651), 475–478. https://doi.org/10.1038/nature22069

The IMBIE team. (2018). Mass balance of the Antarctic ice sheet from 1992 to 2017. *Nature*. https://doi.org/10.1098/rsta.2006.1792

Thomas, R. H., & Bentley, C. R. (1978). A model for Holocene retreat of the West Antarctic Ice Sheet. *Quaternary Research*, *10*(2), 150–170. https://doi.org/10.1016/0033-5894(78)90098-4

Thornalley, D. J. R., Oppo, D. W., Ortega, P., Robson, J. I., Brierley, C. M., Davis, R., … Keigwin, L. D. (2018). Anomalously weak Labrador Sea convection and Atlantic overturning during the past 150 years. *Nature*, *556*(7700), 227–230. https://doi.org/10.1038/s41586-018-0007-4

Thornton, B. F., Geibel, M. C., Crill, P. M., Humborg, C., & Mörth, C.-M. (2016). Methane fluxes from the sea to the atmosphere across the Siberian shelf seas. *Geophysical Research Letters*, *43*(11), 5869–5877. https://doi.org/https://doi.org/10.1002/2016GL068977

Thornton, B. F., Prytherch, J., Andersson, K., Brooks, I. M., Salisbury, D., Tjernström, M., & Crill, P. M. (2020). Shipborne eddy covariance observations of methane fluxes constrain Arctic sea emissions. *Science Advances*, *6*(5), eaay7934. https://doi.org/10.1126/sciadv.aay7934

Tietsche, S., Notz, D., Jungclaus, J. H., & Marotzke, J. (2011). Recovery mechanisms of Arctic summer sea ice. *Geophysical Research Letters*, *38*(2), 1–4. https://doi.org/10.1029/2010GL045698

Timmermans, M.-L., Ladd, C., & Wood, K. (2017). Sea Surface Temperautre. In J. Richter-Menge, J. E. Overland, J. T. Mathis, & E. Osborne (Eds.), *Arctic Report Card 2017*.

Tohjima, Y., Zeng, J., Shirai, T., Niwa, Y., Ishidoya, S., Taketani, F., … Morimoto, S. (2020). Estimation of CH4 emissions from the East Siberian Arctic Shelf based on atmospheric observations aboard the R/V Mirai during fall cruises from 2012 to 2017. *Polar Science*, (xxxx), 100571. https://doi.org/10.1016/j.polar.2020.100571

Topal, D., Ding, Q., Mitchell, J., Baxter, I., Herein, M., Haszpra, T., … Li, Q. (2020). An Internal Atmospheric Process Determining Summertime Arctic Sea Ice Melting in the Next Three Decades: Lessons Learned from Five Large Ensembles and Multiple CMIP5 Climate Simulations. *Journal of Climate*, *33*(17), 7431–7454. https://doi.org/10.1175/JCLI-D-19-0803.1

Turetsky, M., Abbott, B., Jones, M., Walter Anthony, K. M., Olefeldt, D., Schuur, E. A. G., … McGuire, A. D. (2020). Carbon release through abrupt permafrost thaw. *Nature Geoscience*, *Accepted*(February). https://doi.org/10.1038/s41561-019-0526-0

Turetsky, M. R., Jones, M. C., Walter Anthony, K., Olefeldt, D., Schuur, E. A. G., Koven, C., … Grosse, G. (2019). Permafrost collapse is accelerating carbon release. *Nature*, *569*, 32–24.

Turner, A. G., & Annamalai, H. (2012, August). Climate change and the South Asian summer monsoon. *Nature Climate Change*, Vol. 2, pp. 587–595. https://doi.org/10.1038/nclimate1495

Turner, M. G., Calder, W. J., Cumming, G. S., Hughes, T. P., Jentsch, A., LaDeau, S. L., … Carpenter, S. R. (2020). Climate change, ecosystems and abrupt change: science priorities. *Philosophical Transactions of the Royal Society B: Biological Sciences*, *375*(1794), 20190105. https://doi.org/10.1098/rstb.2019.0105

US Forest Service. (2019). *Forest Health Conditions in Alaska - 2019*. Retrieved from http://forestry.alaska.gov/pdfs/2003_fhp_CONDRPT_web_final.pdf

Valentine, D. L., Blanton, D. C., Reeburgh, W. S., & Kastner, M. (2001). Water column methane oxidation adjacent to an area of active hydrate dissociation, Eel River Basin. *Geochimica et Cosmochimica Acta*, *65*(16), 2633–2640. https://doi.org/10.1016/S0016-7037(01)00625-1

Van Den Broeke, M. R., Enderlin, E. M., Howat, I. M., Kuipers Munneke, P., Noël, B. P. Y., Jan Van De Berg, W., … Wouters, B. (2016). On the recent contribution of the Greenland ice sheet to sea level change. *Cryosphere*, *10*(5), 1933–1946. https://doi.org/10.5194/tc-10-1933-2016

van der Zande, R. M., Achlatis, M., Bender-Champ, D., Kubicek, A., Dove, S., & Hoegh-Guldberg, O. (2020). Paradise lost: End-of-century warming and acidification under business-as-usual emissions have severe consequences for symbiotic corals. *Global Change Biology*, *26*(4), 2203–2219. https://doi.org/https://doi.org/10.1111/gcb.14998

Velicogna, I., Sutterley, T. C., & Van Den Broeke, M. R. (2014). Regional acceleration in ice mass loss from Greenland and Antarctica using GRACE time-variable gravity data. *Geophysical Research Letters*, *41*(22), 8130–8137. https://doi.org/10.1002/2014GL061052

Veraverbeke, S., Rogers, B. M., Goulden, M. L., Jandt, R. R., Miller, C. E., Wiggins, E. B., & Randerson, J. T. (2017).

Lightning as a major driver of recent large fire years in North American boreal forests. *Nature Climate Change*, *7*(7), 529–534. https://doi.org/10.1038/nclimate3329

Veron, J. E. N., Hoegh-Guldberg, O., Lenton, T. M., Lough, J. M., Obura, D. O., Pearce-Kelly, P., … Rogers, A. D. (2009). The coral reef crisis: The critical importance of <350 ppm CO2. *Marine Pollution Bulletin*, *58*(10), 1428–1436. https://doi.org/10.1016/j.marpolbul.2009.09.009

Victoria, R. L., Martinelli, L. A., Moraes, J. M., Ballester, M. V., Krusche, A. V., Pellegrino, G., … Richey, J. E. (1998). Surface Air Temperature Variations in the Amazon Region and Its Borders during This Century. *Journal of Climate*, *11*(5), 1105–1110. https://doi.org/10.1175/1520-0442(1998)011<1105:satvit>2.0.co;2

Voigt, C., Marushchak, M. E., Abbott, B. W., Biasi, C., Elberling, B., Siciliano, S. D., … Martikainen, P. J. (2020). Nitrous oxide emissions from permafrost-affected soils. *Nature Reviews Earth & Environment*, *1*(8), 420–434. https://doi.org/10.1038/s43017-020-0063-9

Vonk, J. E., Mann, P. J., Davydov, S., Davydova, A., Spencer, R. G. M., Schade, J., … Holmes, R. M. (2013). High biolability of ancient permafrost carbon upon thaw. *Geophysical Research Letters*, *40*(11), 2689–2693. https://doi.org/10.1002/grl.50348

Wadhams, P. (2012). Arctic ice cover, ice thickness and tipping points. *Ambio*, *41*(1), 23–33. https://doi.org/10.1007/s13280-011-0222-9

Wagner, F H, Hérault, B., Bonal, D., Stahl, C., Anderson, L. O., Baker, T. R., … Aragão, L. E. O. C. (2016). Climate seasonality limits leaf carbon assimilation and wood productivity in tropical forests. *Biogeosciences*, *13*(8), 2537–2562. https://doi.org/10.5194/bg-13-2537-2016

Wagner, Fabien Hubert, Hérault, B., Rossi, V., Hilker, T., Maeda, E. E., Sanchez, A., … Aragão, L. E. O. C. (2017). Climate drivers of the Amazon forest greening. *PLOS ONE*, *12*(7), e0180932. Retrieved from https://doi.org/10.1371/journal.pone.0180932

Wagner, T. J. W., & Eisenman, I. (2015). How Climate Model Complexity Influences Sea Ice Stability. *Journal of Climate*, *28*(10), 3998–4014. https://doi.org/10.1175/jcli-d-14-00654.1

Walker, A. P., Hanson, P. J., De Kauwe, M. G., Medlyn, B. E., Zaehle, S., Asao, S., … Norby, R. J. (2014). Comprehensive ecosystem model-data synthesis using multiple data sets at two temperate forest free-air CO2 enrichment experiments: Model performance at ambient CO2 concentration. *Journal of Geophysical Research: Biogeosciences*, *119*(5), 937–964. https://doi.org/10.1002/2013JG002553

Walker, X. J., Baltzer, J. L., Cumming, S. G., Day, N. J., Ebert, C., Goetz, S., … Mack, M. C. (2019). Increasing wildfires threaten historic carbon sink of boreal forest soils. *Nature*, *572*(7770), 520–523. https://doi.org/10.1038/s41586-019-1474-y

Wallmann, K., Riedel, M., Hong, W. L., Patton, H., Hubbard, A., Pape, T., … Bohrmann, G. (2018). Gas hydrate dissociation off Svalbard induced by isostatic rebound rather than global warming. *Nature Communications*, *9*(1), 83. https://doi.org/10.1038/s41467-017-02550-9

Walsh, J. E., Overland, J. E., Groisman, P. Y., & Rudolf, B. (2011). Ongoing climate change in the arctic. *Ambio*, *40*(SUPPL. 1), 6–16. https://doi.org/10.1007/s13280-011-0211-z

Walsh, M. G., De Smalen, A. W., & Mor, S. M. (2018). Climatic influence on anthrax suitability in warming northern latitudes. *Scientific Reports*, *8*(1). https://doi.org/10.1038/s41598-018-27604-w

Walter Anthony, K., Schneider von Deimling, T., Nitze, I., Frolking, S., Emond, A., Daanen, R., … Grosse, G. (2018). 21st-century modeled permafrost carbon emissions accelerated by abrupt thaw beneath lakes. *Nature Communications*, *9*(1), 3262. https://doi.org/10.1038/s41467-018-05738-9

Wang, Bin, Jin, C., & Liu, J. (2020). Understanding Future Change of Global Monsoons Projected by CMIP6 Models. *Journal of Climate*, *33*(15), 6471–6489. https://doi.org/10.1175/JCLI-D-19-0993.1

Wang, Binbin, Jun, I., Socolofsky, S. A., DiMarco, S. F., & Kessler, J. D. (2020). Dynamics of Gas Bubbles From a Submarine Hydrocarbon Seep Within the Hydrate Stability Zone. *Geophysical Research Letters*, *47*(18), e2020GL089256. https://doi.org/https://doi.org/10.1029/2020GL089256

Wang, J. A., Baccini, A., Farina, M., Randerson, J. T., & Friedl, M. A. (2021). Disturbance suppresses the aboveground carbon sink in North American boreal forests. *Nature Climate Change*, *11*(5), 435–441. https://doi.org/10.1038/s41558-021-01027-4

Wang, J. A., Sulla-Menashe, D., Woodcock, C. E., Sonnentag, O., Keeling, R. F., & Friedl, M. A. (2019). Extensive land cover change across Arctic–Boreal Northwestern North America from disturbance and climate forcing. *Global Change Biology*, (July), 1–16. https://doi.org/10.1111/gcb.14804

Wang, Y., Cheng, H., Edwards, R. L., Kong, X., Shao, X., Chen, S., … An, Z. (2008). Millennial- and orbital-scale changes in the East Asian monsoon over the past 224,000 years. *Nature*, *451*(7182), 1090–1093. https://doi.org/10.1038/nature06692

Way, R. G., Lewkowicz, A. G., & Zhang, Y. (2018). Characteristics and fate of isolated permafrost patches in coastal Labrador, Canada. *Cryosphere*, *12*(8), 2667–2688. https://doi.org/10.5194/tc-12-2667-2018

Weatherdon, L. V., Magnan, A. K., Rogers, A. D., Sumaila, U. R., & Cheung, W. W. L. (2016). Observed and projected impacts of climate change on marine fisheries, aquaculture, coastal tourism, and human health: An update. *Frontiers in Marine Science*, *3*(APR). https://doi.org/10.3389/fmars.2016.00048

Webster, P. J. (2004). *The Elementary Hadley Circulation BT - The Hadley Circulation: Present, Past and Future* (H. F. Diaz & R. S. Bradley, Eds.). https://doi.org/10.1007/978-1-4020-2944-8_2

Weertman, J. (1974). Stability of the Junction of an Ice Sheet and an Ice Shelf. *Journal of Glaciology*, *13*(67), 3–11. https://doi.org/10.3189/s0022143000023327

Weijer, W., Cheng, W., Drijfhout, S. S., Fedorov, A. V, Hu, A., Jackson, L. C., … Zhang, J. (2019). Stability of the Atlantic Meridional Overturning Circulation: A Review and Synthesis. *Journal of Geophysical Research: Oceans*, *124*(8), 5336–5375. https://doi.org/https://doi.org/10.1029/2019JC015083

Weijer, W., Cheng, W., Garuba, O. A., Hu, A., & Nadiga, B. T. (2020). CMIP6 Models Predict Significant 21st Century

Decline of the Atlantic Meridional Overturning Circulation. *Geophysical Research Letters*, *47*(12), e2019GL086075. https://doi.org/https://doi.org/10.1029/2019GL086075

Wernberg, T., Bennett, S., Babcock, R. C., Bettignies, T. De, Cure, K., Depczynski, M., … Tuya, F. (2015). Climate-driven regime shift of a temperate Marine Ecosystem. *Science*, *353*(6295), 169–172. https://doi.org/10.1126/science.aad8745

Wheeler, M. C., & McBride, J. L. (2005). *Australian-Indonesian monsoon BT - Intraseasonal Variability in the Atmosphere-Ocean Climate System* (W. K. M. Lau & D. E. Waliser, Eds.). https://doi.org/10.1007/3-540-27250-X_5

Wilcox, E. J., Keim, D., de Jong, T., Walker, B., Sonnentag, O., Sniderhan, A. E., … Marsh, P. (2019). Tundra shrub expansion may amplify permafrost thaw by advancing snowmelt timing. *Arctic Science*, *5*(4), 202–217. https://doi.org/10.1139/as-2018-0028

Wilkerson, J., Dobosy, R., Sayres, D. S., Healy, C., Dumas, E., Baker, B., & Anderson, J. G. (2019). Permafrost nitrous oxide emissions observed on a landscape scale using the airborne eddy-covariance method. *Atmospheric Chemistry and Physics*, *19*(7), 4257–4268. https://doi.org/10.5194/acp-19-4257-2019

Willis, J. K. (2010). Can in situ floats and satellite altimeters detect long-term changes in Atlantic Ocean overturning? *Geophysical Research Letters*, *37*(6). https://doi.org/https://doi.org/10.1029/2010GL042372

Winton, M. (2006). Does the Arctic sea ice have a tipping point? *Geophysical Research Letters*, *33*(23). https://doi.org/10.1029/2006GL028017

Winton, M. (2011). Do climate models underestimate the sensitivity of Northern Hemisphere sea ice cover? *Journal of Climate*, *24*(15), 3924–3934. https://doi.org/10.1175/2011JCLI4146.1

Wisser, D., Marchenko, S., Talbot, J., Treat, C., & Frolking, S. (2011). Soil temperature response to 21st century global warming: The role of and some implications for peat carbon in thawing permafrost soils in North America. *Earth System Dynamics*, *2*(1), 121–138. https://doi.org/10.5194/esd-2-121-2011

Wood, R. (2012). Stratocumulus clouds. *Monthly Weather Review*, *140*(8), 2373–2423. https://doi.org/10.1175/MWR-D-11-00121.1

Wotton, B. M., Flannigan, M. D., & Marshall, G. A. (2017). Potential climate change impacts on fire intensity and key wildfire suppression thresholds in Canada. *Environmental Research Letters*, *12*(9). https://doi.org/10.1088/1748-9326/aa7e6e

Wouters, B., Martin-Español, A., Helm, V., Flament, T., Van Wessem, J. M., Ligtenberg, S. R. M., … Bamber, J. L. (2015). Dynamic thinning of glaciers on the Southern Antarctic Peninsula. *Science*, *348*(6237), 899–903. https://doi.org/10.1126/science.aaa5727

Wright, J. S., Fu, R., Worden, J. R., Chakraborty, S., Clinton, N. E., Risi, C., … Yin, L. (2017). Rainforest-initiated wet season onset over the southern Amazon. *Proceedings of the National Academy of Sciences of the United States of America*, *114*(32), 8481–8486. https://doi.org/10.1073/pnas.1621516114

Wu, C., Venevsky, S., Sitch, S., Yang, Y., Wang, M., Wang, L., & Gao, Y. (2017). Present-day and future contribution of climate and fires to vegetation composition in the boreal forest of China. *Ecosphere*, *8*(8), e01917. https://doi.org/10.1002/ecs2.1917

Wunderling, N, Donges, J. F., Kurths, J., & Winkelmann, R. (2021). Interacting tipping elements increase risk of climate domino effects under global warming. *Earth System Dynamics*, *12*(2), 601–619. https://doi.org/10.5194/esd-12-601-2021

Wunderling, Nico, Willeit, M., Donges, J. F., & Winkelmann, R. (2020). Global warming due to loss of large ice masses and Arctic summer sea ice. *Nature Communications*, *11*(1), 5177. https://doi.org/10.1038/s41467-020-18934-3

Wunsch, C. (2005). The total meridional heat flux and its oceanic and atmospheric partition. *Journal of Climate*, *18*(21), 4374–4380. https://doi.org/10.1175/JCLI3539.1

Wyatt, A. S. J., Leichter, J. J., Toth, L. T., Miyajima, T., Aronson, R. B., & Nagata, T. (2020). Heat accumulation on coral reefs mitigated by internal waves. *Nature Geoscience*, *13*(1), 28–34. https://doi.org/10.1038/s41561-019-0486-4

Xie, S.-P., & Saiki, N. (1999). Abrupt Onset and Slow Seasonal Evolution of Summer Monsoon in an Idealized GCM Simulation. *Journal of the Meteorological Society of Japan. Ser. II*, *77*(4), 949–968. https://doi.org/10.2151/jmsj1965.77.4_949

Yanai, M. (1964). Formation of tropical cyclones. *Reviews of Geophysics*, *2*(2), 367–414. https://doi.org/https://doi.org/10.1029/RG002i002p00367

Yeager, S. G., Karspeck, A. R., & Danabasoglu, G. (2015). Predicted slowdown in the rate of Atlantic sea ice loss. *Geophysical Research Letters*, *42*(24), 10,704-10,713. https://doi.org/10.1002/2015GL065364

Yin, J., & Goddard, P. B. (2013). Oceanic control of sea level rise patterns along the East Coast of the United States. *Geophysical Research Letters*, *40*(20), 5514–5520. https://doi.org/https://doi.org/10.1002/2013GL057992

Yin, J., Schlesinger, M. E., & Stouffer, R. J. (2009). Model projections of rapid sea-level rise on the northeast coast of the United States. *Nature Geoscience*, *2*(4), 262–266. https://doi.org/10.1038/ngeo462

Zatsepina, O. Y., & Buffett, B. A. (1998). Thermodynamic conditions for the stability of gas hydrate in the seafloor. *Journal of Geophysical Research: Solid Earth*, *103*(B10), 24127–24139. https://doi.org/10.1029/98jb02137

Zeebe, R. E., Ridgwell, A., & Zachos, J. C. (2016). Anthropogenic carbon release rate unprecedented during the past 66 million years. *Nature Geoscience*, *9*(4), 325–329. https://doi.org/10.1038/ngeo2681

Zeng, N., Dickinson, R. E., & Zeng, X. (1996). Climatic impact of Amazon deforestation - A mechanistic model study. *Journal of Climate*, Vol. 9, pp. 859–883. https://doi.org/10.1175/1520-0442(1996)009<0859:CIOADM>2.0.CO;2

Zhao, J., & Johns, W. (2014). Wind-forced interannual variability of the Atlantic Meridional Overturning Circulation at 26.5°N. *Journal of Geophysical Research: Oceans*, *119*(4), 2403–2419. https://doi.org/https://doi.org/10.1002/2013JC009407

Zickfeld, K., B. Knopf V, . Petoukhov, & H. J. Schellnhuber. (2005). Is the Indian summer monsoon stable against global change? *Geophysical Research Letters*, *32*(15), L15707. https://doi.org/10.1029/2005GL022771

Zickfeld, Kirsten, Solomon, S., & Gilford, D. M. (2017). Centuries of thermal sea-level rise due to anthropogenic emissions of short-lived greenhouse gases. *Proceedings of the National Academy of Sciences of the United States of America*, *114*(4), 657–662. https://doi.org/10.1073/pnas.1612066114

Zolkos, S., & Tank, S. E. (2020). Experimental Evidence That Permafrost Thaw History and Mineral Composition Shape Abiotic Carbon Cycling in Thermokarst-Affected Stream Networks. *Frontiers in Earth Science*, *8*, 152. https://doi.org/10.3389/feart.2020.00152

Zou, D., Zhao, L., Sheng, Y., Chen, J., Hu, G., Wu, T., … Cheng, G. (2017). A new map of permafrost distribution on the Tibetan Plateau. *The Cryosphere*, *11*(6), 2527–2542. https://doi.org/10.5194/tc-11-2527-2017

Zou, S., Lozier, M. S., Li, F., Abernathey, R., & Jackson, L. (2020). Density-compensated overturning in the Labrador Sea. *Nature Geoscience*, *13*(2), 121–126. https://doi.org/10.1038/s41561-019-0517-1