

## Supporting Information-1

# Multiple climate tipping points metrics for improved sustainability assessment of products and services

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## S1. Supplementary methods

This section includes (i) review and selection of climate tipping elements, (ii) methods for calculation of *impact of the emission*, capacity and effects from crossing tipping points (including justifications when effects from tipping were not modelled), (iii) reasoning for setting a capacity cutoff, (iv) an example of MCTP calculation with two tipping points, (v) details of the two illustrative simulations presented in the main article and (vi) equations for calculation of impact scores with MCTP and the other metrics used in the case study.

### S1.1 Review and selection of climate tipping elements

The following three criteria were used to select relevant tipping elements in this study:

*Criterion 1: There is an evidence of a critical threshold beyond which a small change in one variable controlling the system (control variable) causes a large qualitative change in the system (that is, the system exhibits threshold behavior).* The change may be abrupt and occur immediately after the cause or gradual and spanning over longer timescales. This criterion excludes those tipping elements for which the transition to a new state is a continuous process without strong nonlinearity or threshold behavior. To ensure that there is minimum understanding of the tipping dynamics supported by scientific evidence we considered only those tipping points, which are described in at least two studies published by different groups in peer-reviewed journals. In cases where there were more studies but the evidence of threshold behavior was debated between them, we conservatively assumed that the potential tipping element does exhibit threshold behavior, thus meeting the criterion.

*Criterion 2: There is an evidence that the system's critical control variable that may pass a threshold is influenced by changes in atmospheric CO<sub>2</sub>-equivalent GHGs concentration.* This influence is typically indirect, meaning that a change in GHG concentrations affects certain parameters that in turn have an influence on the control variable of the system (e.g. GHG concentrations affect atmospheric temperature, which, in turn, influences the freshwater and heat inputs that control the thermohaline circulation in the North Atlantic). This criterion is chosen because climate tipping characterization factors (CF) are for greenhouse gas emissions. Thus, those tipping points that cannot be related to GHG concentrations, but are influenced by e.g. aerosol pollution, are not relevant and are therefore excluded.

*Criterion 3: Tipping threshold estimates and their relative uncertainties can be expressed as global mean temperature change above pre-industrial levels.* Despite relatively uncertain link between global mean temperature and the actual control variable of a tipping element<sup>1</sup>, global mean temperature change is used here as indicator of threshold levels because it was found to be the most

common way of reporting critical thresholds among studies (see references in Table S1). Thus, we excluded tipping elements for which threshold estimations are only given based on local physical parameters, e.g. sea ice extent/thickness, rate of ocean current or precipitation rate, for which the calculation of the corresponding level of atmospheric GHGs concentration equivalents (i.e. the concentration necessary to calculate the remaining capacity) is either not possible or highly uncertain.

An overview of all potential tipping elements is given in Table S1. Table S2 provides an overview of potential occurrence of the 13 selected tipping elements under the considered Representative Concentration Pathways (RCP).

82 **Table S1:** Overview of potential tipping elements from the literature and their fulfilment of the selection criteria (1, 2 and 3) in this study. Selected tipping  
83 points that meet all criteria are highlighted in bold. When available, ranking based on likelihood is shown for illustrative purposes even though it was not used  
84 as a criterion. Fulfilment of selection criteria is indicated with ‘y’ = yes; ‘n’ = no; ‘un’ = uncertain.

Tipping element	Critical tipping threshold <sup>a</sup> (global mean temperature above pre-industrial in °C, unless differently indicated)		Climatic triggering and main feedback mechanism	Transition period <sup>b</sup> (yr)		Likelihood		Fulfilled selection criteria		
		Reference			Reference	Relative likelihood <sup>2</sup>	Likelihood with increasing global warming <sup>3</sup>	1	2	3
Arctic summer sea ice loss (AS)	1.2 – 2.7 <sup>c</sup>	<sup>4</sup>	Melting of sea exposes larger areas of ocean surface to solar radiation, decreasing the albedo and increasing heat absorption by the ocean, thus amplify the warming. This loss of sea ice could lead to ice cap melting beyond certain size/thickness at which complete melting is likely to occur every summer.	10	<sup>4</sup>	1	1	y	y	y
	1 – 3	<sup>5</sup>								
	1.5 – 2 <sup>d</sup>	<sup>6, 7</sup>								
	2.2 – 2.7 (2.5)	<sup>8</sup>								
Greenland ice sheet melt (GI)	1.7 – 2.7 <sup>c</sup>	<sup>4</sup>	Increased air temperatures cause surface ice melting, which lowers ice altitude and increases surface temperature (due to higher temperatures at lower elevation) causing further warming and melting (melt-elevation feedback) to a point beyond which there is net mass loss and GI shrinks radically <sup>9</sup> .	300 – 7500 (1500) <sup>c</sup>	<sup>10</sup>	2	2	y	y	y
	1 – 3	<sup>5</sup>								
	0.8 – 3.2 (1.6) <sup>e</sup>	<sup>11</sup>								
	1.9 – 5.1 (3.1) <sup>e</sup>	<sup>12</sup>								
	2 – 3	<sup>13</sup>								
	2 – 4	<sup>14</sup>								
West Antarctic ice sheet collapse (AI)	3.7 – 5.7 <sup>c</sup>	<sup>4</sup>	The collapse is due to the combination of (i) surface melting (see GI) and (ii) the retreat of the submerged grounding line caused by the intrusion of warmer ocean water, which increases the ice flux and induces further retreat <sup>3,9</sup> .	100 – 2500 (500) <sup>c</sup>	<sup>10</sup>	3	4	y	y	y
	1 – 3	<sup>5</sup>								
	4 <sup>f</sup>	<sup>15</sup>								
	1 – 5.7	<sup>16</sup>								
Amazon rainforest dieback (AF)	3.7 – 4.7 <sup>c</sup>	<sup>4</sup>	Warmer temperatures cause reduction in precipitations, lengthening of the dry season and directly affect vegetation productivity, leading to forest dieback, which in turn further reduces precipitations <sup>17</sup> .	50 – 250 (50) <sup>c</sup>	<sup>10</sup>	3	3	y	y	y
	3 – 5	<sup>5</sup>								
	2.5, 6.2 <sup>g</sup>	<sup>18</sup>								
	2, 3, 4 <sup>h</sup>	<sup>17</sup>								
Sahara/Sahel and West African monsoon shift (AM)	3.7 – 4.7 <sup>c</sup>	<sup>4</sup>	Warming of sea surface temperature influences the direction of the West African monsoon, which in turn affects rainfall in the Sahara/Sahel region. It is uncertain whether WAS will shift northward (leading to increased rainfall) or southward (leading to further drying of the Sahel) <sup>3</sup> .	10	<sup>4</sup>	4	6	y	y	y
	3 – 5	<sup>5</sup>								
	2.1, 2.8, 3.5 <sup>i</sup>	<sup>18</sup>								

<b>Boreal forest dieback (BF)</b>	3.7 – 5.7 <sup>c</sup>	<sup>4</sup>	Increased water stress, peak summer heat stress, vulnerability to disease and fire frequency due to higher temperatures cause boreal forest dieback and transition to open woodlands or grasslands, which in turn would amplify summer heat stress, drying and fire frequency <sup>3</sup> .	50	<sup>4</sup>	5	4	y	y	y
	3 – 5	<sup>5</sup>								
<b>El Niño-Southern Oscillation change in amplitude (EN)</b>	3.7 – 6.7 <sup>c</sup>	<sup>4</sup>	Increased heat uptake in the equatorial Pacific could lead to a permanent deepening of the thermocline, which could result in more persistent El Niño-like conditions. However, it is not excluded that stronger warming of the west Equatorial Pacific than the east could lead to more persistent La Niña-like conditions <sup>4</sup> . Complex and uncertain mechanism.	100	<sup>4</sup>	5	5	y	y	y
	3 – 5	<sup>5</sup>								
<b>Atlantic thermohaline circulation shutoff (TC)</b>	3.7 – 5.7 <sup>c</sup>	<sup>4</sup>	Addition of freshwater in the North Atlantic (due to sea ice and Greenland ice sheet melting, river inputs and ocean precipitation) may reduce the density-driven sinking of North Atlantic waters until the Atlantic thermohaline circulation is significantly slowed down or even stopped <sup>19</sup> .	10 – 250 (50) <sup>c</sup>	<sup>10</sup>	5	5	y	y	y
	3 – 5	<sup>5</sup>								
	1.4, 1.6, 1.9 <sup>i</sup>	<sup>18</sup>								
	6 – 8 <sup>j</sup>	<sup>20</sup>								
	1.4, 1.6, 2.2, 2.5 <sup>l</sup>	<sup>21</sup>								
<b>Permafrost loss (P)</b>	5.6	<sup>18</sup>	Thawing of permafrost (highly rich in organic carbon) in the northeastern Siberia (Yedoma) triggers biochemical decomposition of the organic matter, which generates heat that further increases warming and melting.	<100	<sup>4</sup>	Not assessed	7	y	y	y
	> 5	<sup>5</sup>								
	> 5	<sup>16</sup>								
Indian summer monsoon collapse	3 – 5	<sup>5</sup>	Monsoon dynamics depend on heat and pressure differences between land and ocean. Black carbon and aerosol emissions on land reduce land-absorbed solar radiation and the resulting lower warming of the land compared to the ocean weakens the monsoon to its eventual collapse. On the contrary, increased warming over land due to GHG emissions generally strengthens the monsoon. Increasing global average temperature influences the monsoon but does not lead to its collapse <sup>19</sup> .	1	<sup>4</sup>	Not assessed	Not assessed	y	n	y
Tundra loss	7.2	<sup>18</sup>	Warmer temperatures enable northward expansion of boreal forest in replacement of tundra regions, initiating a positive	100	<sup>4</sup>	Not assessed	Not assessed	y	y	n

			snow-albedo feedback where the darker surface covered with trees reduces snow albedo and amplifies warming.							
Marine methane hydrates release	-	-	Warmer ocean temperatures could melt the large amount of frozen methane hydrates and the gas bubbles they trap beneath sediments in the ocean floor, which would then be released in the atmosphere causing further warming <sup>3</sup> .	1000 – 100,000	<sup>4</sup>	Not assessed	Not assessed	y	y	n
Ocean anoxia	-	-	Warmer ocean temperatures decrease ventilation of deep water and solubility of O <sub>2</sub> in surface water leading to widespread oceanic anoxic conditions. Low oxygen levels in the ocean increment the nitrous oxide emissions <sup>22</sup> and may have other consequences that could reduce ocean's CO <sub>2</sub> absorption capacity <sup>23</sup> .	10,000	<sup>4</sup>	Not assessed	Not assessed	y	y	n
Arctic ozone loss	-	-	“Global warming implies global cooling of the stratosphere that supports formation of ice clouds, which in turn provide a catalyst for stratospheric ozone destruction” <sup>24</sup>	1	<sup>4</sup>	Not assessed	Not assessed	un	y	n
Antarctic bottom water formation collapse	-	-	Increased precipitations at high latitudes resulting from global warming cause surface water freshening around Antarctica, which suppresses ocean convection and so bottom water formation <sup>25</sup>	≈ 100	<sup>4</sup>	Not assessed	Not assessed	un	y	n
<b>Alpine glaciers loss (AG)</b>	2	<sup>26</sup>	Increased temperatures cause reduction in snow and ice cover, originating a positive ice-albedo feedback, and prolongation of the melting season, which destabilizes the glacier mass balance towards glacier thinning and disintegration <sup>26</sup>	100	<sup>26</sup>	Not assessed	Not assessed	y	y	y
	1 – 3	<sup>5</sup>								
<b>Coral reefs deterioration (CR)</b>	450 – 500 ppm [CO <sub>2</sub> ] <sub>atm</sub>	<sup>27</sup>	Increased sea temperature due to global warming results in coral bleaching (breakdown of symbiosis between corals and the algae that live inside their tissues) and mortality. Moreover, increased atmospheric CO <sub>2</sub> concentration means higher uptake by oceans, where CO <sub>2</sub> reacts to form carbonic acid, which reduces the availability of carbonate ions and the rate of calcification of corals ultimately favoring erosion. Both processes trigger multiple ecological feedback loops that eventually drive reefs to a non-coral dominated state <sup>27</sup> .	few years to decades (<100)	<sup>28</sup>	Not assessed	Not assessed	y	y	y
	1 – 3	<sup>5</sup>								
	1.25 – 2 <sup>k</sup>	<sup>29</sup>								

East Antarctic ice sheet collapse	>5	<sup>16, 5</sup>	Same dynamics as AI	200 – 800	<sup>30</sup>	Not assessed	Not assessed	un	y	y
Arctic winter sea ice loss (AW)	4.5, 5.2, 6.2, 7.4, 8.2 <sup>g</sup>	<sup>18</sup>	Besides ice-albedo feedback (see Arctic summer sea ice) also reduced ice thickness creates a positive feedback that leads to complete ice loss.	10	<sup>18</sup>	Not assessed	Not assessed	y	y	y
	>5	<sup>5</sup>								
North Atlantic subpolar gyre convection collapse (SG)	1.1, 1.4, 1.6, 1.7, 1.9, 2.0, 3.8 <sup>l</sup>	<sup>21</sup>	Warming and freshening of the North Atlantic subpolar gyre (an area of cyclonic ocean circulation in the Northwest Atlantic) leads to stratification (as consequence of lower surface density), that weakens the local deep convection, which in turn amplifies the stratification (because of reduced inflow of saltier water from the surroundings), eventually leading to permanent convection collapse. This collapse involves only the subpolar gyre (which is part of the TC) and not the whole North Atlantic TC.	10	<sup>18, 21</sup>	Not assessed	Not assessed	y	y	y
West tropical Indian oceanic bloom	10.9	<sup>18</sup>	“This event features an increase in equatorial upwelling, which is due to a general increase in oceanic velocity and divergence at the equator associated with enhanced wind stress at the surface linked to changes in monsoon regime. As a consequence of this increased divergence in the equatorial area, the upwelling increases, bringing a large amount of nutrients to the surface that are then advected toward the coast of Somalia, where the bloom is maximal” <sup>18</sup> .	≈ 10	<sup>18</sup>	Not assessed	Not assessed	n	un	n
Abrupt sea ice increase in Southern Ocean	1.6	<sup>18</sup>	Warming causes deep ocean convection to stop in the Indian sector of the Southern Ocean, which enables the formation of a fresh surface layer and hence of sea ice.	≈ 10	<sup>18</sup>	Not assessed	Not assessed	n	y	n
Abrupt Tibetan snow melt	1.4 – 2.2	<sup>18</sup>	Rising temperature “drives the system into a regime where the annual mass flux balance becomes negative and snow becomes a seasonal phenomenon” leading to a sudden loss of snow <sup>18</sup> .	≈ 10	<sup>18</sup>	Not assessed	Not assessed	n	y	y

85 <sup>a</sup> Estimates are given as either intervals or single data points.

86 <sup>b</sup> Time required for the full effect to unfold.

87 <sup>c</sup> Data from Lenton et al.<sup>4</sup> has been converted from 1980-1999 to 1850-1900 (pre-industrial) reference period by adding the average temperature change difference (0.7°C) between 1850-1900 and 1986-2005 periods found in the literature<sup>14,31</sup> (Berkeley Earth global land and ocean data), assuming that the mean of 1980-1999 and 1986-2005 periods are not significantly different.

88 <sup>d</sup> Projections do not refer to the range of global warming after which every Arctic summer will be ice free, but they represent the range of warming after which there is 30% (under 1.5°C) to 100% (under 2°C) probability of occurrence of at least an ice free summer by 2100. For example, at 2 degrees there is 100% probability that an ice-free summer will occur before 2100, but it is not a yearly recurring event. Thus, the

90 value does not indicate whether a shift to a different state (an ice-free state) has occurred.

91 <sup>e</sup> Best estimate in brackets.

92

93 <sup>f</sup> Above 4°C it is more likely than not that the AI will collapse.

94 <sup>g</sup> Separate values obtained from different climate models under RCP8.5 assumptions.

95 <sup>h</sup> Thresholds above which around 20, 70 and 80% of dieback is inevitable, respectively.

96 <sup>i</sup> Separate values obtained with the same model under different RCP assumptions.

97 <sup>j</sup> Probability of TC collapse increases from 11% at 6°C warming to 30% at 8°C warming.

98 <sup>k</sup> Range at which more than 90% of reef cells are at risk of long-term degradation depending on the thermal tolerance of coral reefs.

99 <sup>l</sup> Separate values obtained from different climate models and emission scenario



**Arctic summer sea ice loss:** Future loss of Arctic summer sea ice has been extensively studied in the literature (see for instance references <sup>32,33,34</sup>). Some authors argue that abrupt transition to summer ice-free conditions is not likely based on the fact that the loss is in principle reversible<sup>35,36</sup>. Others affirm that irreversibility is not a prerequisite for being a tipping point and consider Arctic summer sea ice loss as one of the first tipping points to be triggered as global temperature increases<sup>3,37,38</sup>. Despite this uncertainty, Arctic summer sea ice is conservatively assumed to be a tipping point in this study.

**West Antarctic ice sheet collapse:** Initial estimations place a potential threshold for AI collapse at around 4°C warming or above. However, more recent studies highlight that AI might have already started tipping<sup>39,40</sup> even though no clear threshold range is indicated. Based on this, recent review studies expanded the range with a potential threshold already at 1°C<sup>5,16</sup>. All three criteria are met for this tipping element.

**Permafrost loss:** Initial projections of permafrost melt did not show evidence of a critical threshold, however recent work has suggested that at least one large area of permafrost could exhibit coherent threshold behavior<sup>41–43</sup>. Based on this, criterion 1 is met.

**Indian summer monsoon collapse:** Increasing global average temperature influences the monsoon but does not lead to its collapse<sup>19</sup>, therefore the second criterion is not met.

**Tundra loss:** According to Lenton et al.<sup>4</sup> tundra loss is not considered a tipping point because “the transition from tundra to boreal forest is a continuous process without strong non-linearity or threshold behavior”. However, a more recent study finds abrupt loss of tundra (in terms of roughly 70% boreal forest northward expansion in 100 years) at 7.2°C above pre-industrial, despite it is acknowledged that tundra loss is a gradual transition<sup>18</sup>. Also Scheffer et al.<sup>44</sup> support the fact that “climate change may invoke massive nonlinear shifts in boreal biomes” including tundra loss. Despite the contrasting findings, it is conservatively decided that tundra loss does exhibit threshold behavior and thus meets the first criterion. Threshold estimations in terms of global temperature change above pre-industrial levels were found to be limited to the single value reported by ref. <sup>18</sup>, with no information about uncertainty. Due to this lack of uncertainty estimations, criterion three is not met.

**Marine methane hydrates release:** The release of methane via gas hydrate degradation is considered a ‘slow’ tipping point leading to a long-term chronic release of methane on timescales of millennia and longer<sup>45</sup>. Due the length of the transition, this potential tipping point does not meet the definition of ref. <sup>4</sup> because it is unlikely that qualitative changes in this Earth system will occur within this millennium. Despite this, critical thresholds have been proposed suggesting that there is potential for

threshold behavior<sup>45</sup>, thus meeting the first selection criterion. However, estimations are only based on local parameters such as ocean temperature increase, thus criterion no. 3 is not met. In addition, it is uncertain whether the released methane will actually reach the atmosphere in such amounts as to significantly influence global GHGs concentration. Many biogeochemical sinks and physical processes could prevent much of the gas from reaching the sea-air interface and being injected into the atmosphere<sup>46</sup>, with implications on quantification of effects on the global climate system.

***Ocean anoxia:*** Whether ocean anoxia represents an actual tipping point is still debated. While some believe that anoxia events can lead to major regime shifts in relatively rapid time<sup>47</sup>, others claim that a shift to anoxic state requires too long periods (around 10000 years) for being considered a tipping point<sup>4</sup>. Ocean anoxia is not considered an immediate climate change concern, however it is not excluded that human-induced warming could increase nutrient weathering rates, which could cause ocean anoxia (past ocean anoxia events are thought to be caused by global warming)<sup>48</sup>. Even if considering the phenomenon as a tipping point, criterion three is not met due to lack of threshold estimations.

***Arctic ozone loss:*** It is currently unclear whether a tipping point exists for Arctic ozone<sup>24,26</sup>. Feedback mechanisms on the climate system due to a large-scale depletion of Arctic ozone are also poorly described in the reviewed literature. According to Baldwin et al.<sup>49</sup>, there are interactions between the climate and the state of the ozone layer, e.g. as ozone is a greenhouse gas, its depletion could cause cooling of the lower stratosphere. Overall, there is lack of evidence supporting that Arctic ozone loss is a tipping element and no threshold estimations are reported (criterion 1 and 2 not met).

***Antarctic bottom water formation collapse:*** There is evidence that bottom water formation decreases under climate change scenario simulations<sup>25,50–52</sup>, however it is still not clear whether the phenomenon has threshold behavior. As no threshold estimations were found, this potential tipping element does not meet criterion 3.

***Alpine glaciers loss:*** Different model simulations of Alpine glacier extent demonstrate that an increase in global mean air temperature of 2°C leads to an almost complete loss of glaciers in the Alps<sup>53–55</sup>. Thus, Alpine glacier loss is selected as tipping element in our study.

***Coral reefs deterioration:*** There is agreement that rapid climate change and ocean acidification could lead coral to the functional collapse of coral reefs<sup>16,27–29,56</sup>. Despite this, some argue that it is still unclear whether there is a large-scale tipping point<sup>41</sup>. Due to this uncertainty, coral reefs deterioration is conservatively assumed to be a tipping element.

**East Antarctic ice sheet collapse:** Not enough evidence was found to conclude that the collapse of East Antarctic ice sheet could show threshold behavior. More research is needed to understand and quantify the potential as a major tipping element in the Earth's climate system<sup>57</sup>.

**Arctic winter sea ice loss:** According to Kopp et al.<sup>58</sup>, “the evidence that winter Arctic sea ice is a tipping element is stronger than for summer Arctic sea ice” and other authors found abrupt year-round ice loss in their simulations<sup>18</sup> (meeting criterion 1).

**North Atlantic Subpolar Gyre convection collapse:** Some models forecast a collapse of the SG, where the deep convection in the Labrador sea shuts off in response to climatic conditions<sup>59–61</sup>. In addition, two recent studies have identified the potential existence of a tipping point for the collapse of the SG<sup>18,21</sup>. All three criteria are met for this element.

**West tropical Indian oceanic bloom; Abrupt sea ice increase in Southern Ocean; Abrupt Tibetan snow melt:** these potential abrupt events are described by only one of the reviewed studies<sup>18</sup>, hence the first selection criterion is not met. In addition, no uncertainty of threshold estimations are found for the first two candidates.

**Table S2:** Overview of selected tipping elements with relative temperature threshold intervals and their occurrence under the chosen RCP pathways.

Selected tipping element	Temperature threshold range <sup>‡</sup> (global mean temperature above pre-industrial level in °C)	Occurrence		
		RCP4.5	RCP6	RCP8.5
Arctic summer sea ice loss (AS)	1.5 – 2.6	Expected	Expected	Expected
Greenland ice sheet melt (GI)	1.6 – 3.5	Potential	Expected	Expected
West Antarctic ice sheet collapse (AI)	1.9 – 4.8	Potential	Potential	Expected
Amazon rainforest dieback (AF)	2.8 – 5.0	Potential	Potential	Expected
Boreal forest dieback (BF)	3.4 – 5.4	Not expected	Potential	Expected
El Niño-Southern Oscillation change in amplitude (EN)	3.4 – 5.9	Not expected	Potential	Expected
Permafrost loss (P)	5 – 8.5 <sup>‡</sup>	Not expected	Not expected	Expected
Arctic winter sea ice loss (AW)	4.8 – 8.2	Not expected	Not expected	Expected
Atlantic thermohaline circulation shutoff (TC)	3.1 – 4.6	Not expected	Potential	Expected

North Atlantic subpolar gyre convection collapse (SG)	1.2 <sup>§</sup> – 3.8	Potential	Expected	Expected
Sahara/Sahel and West African monsoon shift (AM)	2.9 – 4.4	Potential	Potential	Expected
Alpine glaciers loss (AG)	1.2 <sup>§</sup> – 3.0	Expected	Expected	Expected
Coral reefs deterioration (CR)	1.2 <sup>§</sup> – 2.5	Expected	Expected	Expected

<sup>†</sup> Assigned by taking the mean of the lower and upper bounds of available intervals following the approach in ref. <sup>62</sup>, if not differently specified.

<sup>§</sup> Literature data reports lower bounds (see Table S1), however 1.2°C was chosen arbitrarily to exclude the possibility that a tipping has already been crossed.

<sup>‡</sup> As no specific upper bound is found (see Table S1), 8.5°C was chosen because it corresponds to the maximum temperature possibly reachable under the selected RCP pathways within year 2500.

## S1.2 Calculation of the *impact of the emission*, $I_{\text{emission},i,j}(T_{\text{emission}})$

Recall, that the *impact of the emission* for a chosen tipping point is calculated following the approach in ref. <sup>63</sup>, and is renamed to the absolute climate tipping potential of gas  $i$  integrated between the emission year  $T_{\text{emission}}$  and the year of tipping  $T_{\text{tipping},j}$  ( $ACTP_{i,j}(T_{\text{emission}})$ ) divided by the radiative efficiency of 1 ppm CO<sub>2</sub> ( $RE_{\text{CO}_2}$ ):

$$I_{\text{emission},i,j}(T_{\text{emission}}) = \frac{ACTP_{i,j}(T_{\text{emission}})}{RE_{\text{CO}_2}} = \frac{\sum_{k=1}^n RF_i(T_{k-1}) \cdot \Delta T}{RE_{\text{CO}_2}} \quad (S1)$$

Where  $j$  indicates the  $j$ th tipping point occurring after the emission year and  $RF_i$  is the radiative forcing of gas  $i$ . The radiative forcing is usually expressed (as a function of continuous time  $t$ ) as the product of the radiative efficiency of gas  $i$  ( $A_i$ ) and the impulse response function ( $IRF$ ), which for most non-CO<sub>2</sub> GHGs is represented with a single exponential decay (eq. S2) and for CO<sub>2</sub> with a sum of  $l$  exponentials (eq. S3)<sup>64</sup>:

$$RF_i(t) = A_i \cdot IRF_i(t) = A_i \left[ e^{-\frac{t}{\tau_i}} \right] \quad (S2)$$

$$RF_{\text{CO}_2}(t) = A_{\text{CO}_2} \cdot IRF_{\text{CO}_2}(t) = A_{\text{CO}_2} \left[ a_0 + \sum_l a_l e^{-\frac{t}{\tau_l}} \right] \quad (S3)$$

The  $IRF$  describes the decay with continuous (and relative) time  $t$  of a perturbation in atmospheric concentration of gas  $i$  after a pulse emission considering how quick the substance is removed from the atmosphere. A summary of the different terms and parameters for calculation of the *impact of the emission* is found in Tables S3, S4 and S5.

Radiative forcing function which was expressed as Riemann sum in eq. S1, can also be solved analytically (eq. S4 for a non-CO<sub>2</sub> gas  $i$  and eq. S5 for CO<sub>2</sub>):

$$I_{\text{emission},i,j}(T_{\text{emission}}) = \frac{A_i \tau_i \left[ 1 - e^{-\frac{(T_{\text{tipping},j} - T_{\text{emission}})}{\tau_i}} \right]}{RE_{\text{CO}_2}} \quad (\text{S4})$$

$$I_{\text{emission},\text{CO}_2,j}(T_{\text{emission}}) = \frac{A_{\text{CO}_2} \left[ a_0 (T_{\text{tipping},j} - T_{\text{emission}}) + \sum_l a_l \tau_l \left( 1 - e^{-\frac{(T_{\text{tipping},j} - T_{\text{emission}})}{\tau_l}} \right) \right]}{RE_{\text{CO}_2}} \quad (\text{S5})$$

As can be seen, for gas  $i$  the *impact of the emissions* occurring at different times is different as it depends on the distance between the emission year and the year of tipping.

**Table S3:** Overview of terms used in the calculation of  $I_{\text{emission},i,j}(T_{\text{emission}})$ .

Parameter	Name	Unit	Definition	Note
$I_{\text{emission},i,j}(T_{\text{emission}})$	Impact of the emission of gas $i$ emitted at $T_{\text{emission}}$ for the $j$ th tipping point	ppm CO <sub>2</sub> e · yr · kg <sub>i</sub> <sup>-1</sup>	Time-integrated change in atmospheric CO <sub>2</sub> -equivalent concentration due to a pulse emission of gas $i$ from $T_{\text{emission}}$ to $T_{\text{tipping},j}$	
$ACTP_{i,j}(T_{\text{emission}})$	Absolute Climate Tipping Potential of gas $i$	W · m <sup>-2</sup> · yr · kg <sub>i</sub> <sup>-1</sup>	Time-integrated radiative forcing due to a pulse emission of gas $i$ from $T_{\text{emission}}$ to $T_{\text{tipping},j}$	
$j$	-	-	Index for the sequence of occurring tipping points in a given iteration	An index taking numerical values was preferred rather than a qualifier for the subscript $j$ , because the order of occurrence of tipping points can be different in different simulations
$T_{\text{emission}}$	Emission year ( <i>interval time</i> )	yr	Any year starting from 2021 in which an emission can occur. $T_{\text{emission}}$ indicates a specific time interval of 1 year.	See Table S7
$t$	Generic time ( <i>point in time</i> )	yr	Continuous, point time expressed with real numbers	See Table S7
$n$	-	-	Number of time steps	Equal to the difference between the year

				of tipping $T_{\text{tipping},j}$ (i.e. the year when the $j$ th tipping point is exceeded) and the year of emission, $T_{\text{emission}}$ .
$RF_{i/\text{CO}_2}$	Radiative forcing	$\text{W} \cdot \text{m}^{-2} \cdot \text{kg}_{i/\text{CO}_2}^{-1}$	Radiative forcing due to a pulse emission of gas $i/\text{CO}_2$	
$RE_{\text{CO}_2}$	Radiative efficiency per ppm $\text{CO}_2$	$\text{W} \cdot \text{m}^{-2} \cdot \text{ppm CO}_2^{-1}$	Radiative forcing of 1 ppm $\text{CO}_2$ with a current background concentration of 409 <sup>a</sup> ppm	Reference value: $1.31 \cdot 10^{-2}$ (calculated based on ref. <sup>65</sup> ).
$IRF_{i/\text{CO}_2}$	Impulse Response Function	unitless	Fraction of gas $i/\text{CO}_2$ remaining in the atmosphere at time $t$ after a pulse emission	
$A_{i/\text{CO}_2}$	Radiative efficiency of gas $i/\text{CO}_2$	$\text{W} \cdot \text{m}^{-2} \cdot \text{kg}_{i/\text{CO}_2}^{-1}$	Radiative forcing per unit mass increase in atmospheric abundance of gas $i/\text{CO}_2$	See Table S4.
$\tau_i$	Atmospheric lifetime (or adjustment time)	yr	Time scale characterizing the decay of a pulse emission input into the atmosphere <sup>66</sup>	Different from the lifetime of a gas, as it accounts for the effects of feedbacks resulting from a pulse emission <sup>67</sup> . See Table S4.
$a_l$ $\{l=0,1,2,3\}$	-	unitless	Weight of each exponential	See Table S5.
$\tau_l$ $\{l=1,2,3\}$	-	yr	Decay times of each exponential	See Table S5.
$l$	-	-	Number of exponentials	Up to 4

<sup>a</sup> Annual average of  $\text{CO}_2$  in situ air measurements (February 2018 - January 2019) from Mauna Loa Observatory, Hawaii<sup>68</sup>.

**Table S4:** Radiative efficiency ( $A_i$ ) and atmospheric lifetime ( $\tau_i$ ) for  $\text{CO}_2$ ,  $\text{CH}_4$  and  $\text{N}_2\text{O}$ <sup>66</sup>.  $A_i$  values are converted from  $\text{W} \cdot \text{m}^{-2} \cdot \text{ppbv}_i^{-1}$  to  $\text{W} \cdot \text{m}^{-2} \cdot \text{kg}_i^{-1}$  using previous methods<sup>64</sup>.

	$A_i (\text{W} \cdot \text{m}^{-2} \cdot \text{kg}_i^{-1})$	$\tau_i (\text{yr})$
$\text{CO}_2$	$1.67\text{E}-15^{\text{a}}$	-
$\text{CH}_4$	$1.82\text{E}-13^{\text{b}}$	12
$\text{N}_2\text{O}$	$3.87\text{E}-13$	114

<sup>a</sup> Obtained using the updated radiative efficiency per ppm of  $\text{CO}_2$  ( $RE_{\text{CO}_2}$ ) presented in Table S3 ( $1.31 \cdot 10^{-2} \text{ W} \cdot \text{m}^{-2} \cdot \text{ppm CO}_2^{-1}$ ).

<sup>b</sup> The calculation includes multiplication by a factor 1.4 to account for indirect radiative effects of methane emissions<sup>66</sup>.

**Table S5:** Constant parameter values for calculation of radiative forcing of CO<sub>2</sub> from ref. <sup>69</sup>.

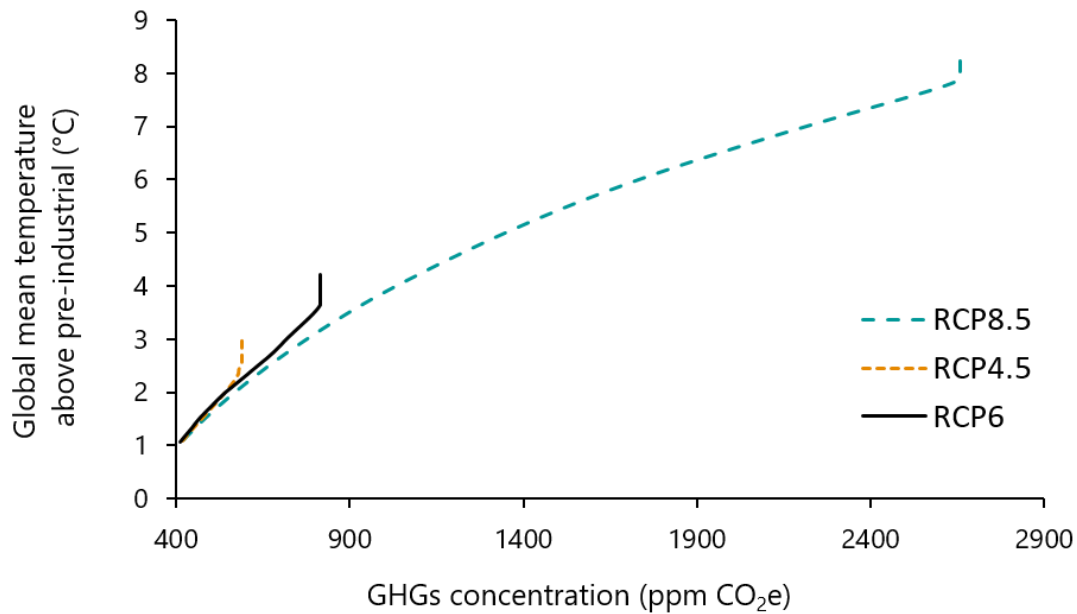
Parameter	1 <sup>st</sup> term ( $l=0$ )	2 <sup>nd</sup> term ( $l=1$ )	3 <sup>rd</sup> term ( $l=2$ )	4 <sup>th</sup> term ( $l=3$ )
$a_l$ (unitless)	2.123E-01 ( $a_0$ )	2.444E-01	3.360E-01	2.073E-01
$\tau_l$ (yr)	-	3.364E+02	2.789E+01	4.055E+00

### S1.3 Calculation of capacity, $CAP_j(T_{\text{emission}})$

Recall, that the  $CAP_j(T_{\text{emission}})$  (ppm CO<sub>2</sub>e · yr) is given:

$$CAP_j(T_{\text{emission}}) = C(T_{\text{tipping},j}) \cdot (T_{\text{tipping},j} - T_{\text{emission}}) - \sum_{k=1}^n [C(T_{k-1}) + C_{\text{tip}}(T_{k-1})] \cdot \Delta T \quad (S6)$$

where,  $C(T_{\text{tipping},j})$  is the atmospheric CO<sub>2</sub>-equivalent concentration at the year of tipping  $T_{\text{tipping},j}$ ,  $C(T)$  is the CO<sub>2</sub>-equivalent concentration from background emissions at time  $T$  and  $C_{\text{tip}}(T)$  is the change in CO<sub>2</sub>-equivalent concentration at time  $T$  caused by all tipping points occurred before  $T_{\text{emission}}$  (all terms expressed in ppm CO<sub>2</sub>e). Determination of  $C(T_{\text{tipping},j})$  and  $C(T)$  depends on the background development of GHG emissions as predicted by the RCP pathways. Differences in the relationship between projected GHGs concentration and the corresponding equilibrium temperature between the pathways (Figure S1) results in different remaining capacity to the same tipping point when calculated with the three RCPs.



**Figure S1:** Relationship between projected GHGs concentration and temperature change in the chosen RCP pathways.

### 266 *S1.3.1 Calculation of effects from tipping, $C_{\text{tip}}(T)$*

267  $C_{\text{tip}}(T_{\text{effect}})$  is defined as the sum of the change in CO<sub>2</sub>-equivalent concentration caused by all tipping  
 268 points occurred before the emission year  $T_{\text{emission}}$ .  $T_{\text{effect}}$  is used here (rather than generic year  $T$ ) to  
 269 represent the year in which effects caused by all these tipping points unfold. The resulting overall  
 270 effect from tipping is therefore expressed as a sum of effects of each tipping element (eq S7):

$$272 \quad C_{\text{tip}}(T_{\text{effect}}) = \sum_a C_{\text{tip},a}(T_{\text{effect}}) \quad (S7)$$

273  
 274 where  $a$  indicates a concrete tipping element that passed its tipping point before  $T_{\text{emission}}$ , and  $C_{\text{tip},a}$   
 275 is the CO<sub>2</sub>-equivalent concentration increase caused by crossing the tipping point of the tipping  
 276 element  $a$ . Here we use subscript  $a$  indicating a specific tipping element (e.g.  $a$  = Arctic summer sea  
 277 ice loss), rather than subscript  $j$  because the effect from tipping depends on the tipping element being  
 278 crossed and not on its order of occurrence.

279 For Arctic summer and winter sea ice loss (AS and AW), the effect from tipping is calculated  
 280 based on the RF change due to reduced sea ice albedo (Table S6), which is then converted to annual  
 281 concentration increase using the radiative efficiency of CO<sub>2</sub> per 1 ppm (found in Table S3). This  
 282 effect is assumed to unfold completely from the year after tipping and to remain constant over the  
 283 years, as the evolution of radiative forcing changes after tipping is unknown. Thus, at any year after  
 284 tipping ( $T_{\text{effect}}$ ),  $C_{\text{tip},AS}$  and  $C_{\text{tip},AW}$  are found to be 22.2 and 52.0 ppm CO<sub>2</sub>e (Figure S2). For  
 285 Greenland ice sheet melt (GI), West Antarctic ice sheet collapse (AI), El Niño-Southern Oscillation  
 286 change in amplitude (EN), Permafrost loss (P), Amazon (AF) and Boreal forest (BF) tipping points, a  
 287 dynamic (i.e. changing over time) effect (in terms of equivalent GHG concentration increase) was  
 288 calculated by adapting the method used in Levasseur et al.<sup>70</sup>. Estimates of carbon emissions that could  
 289 be released after tipping (shown in Table S6) were used to calculate the dynamic effects, considering  
 290 that all emitted carbon is in the form of CO<sub>2</sub>, except for emissions from permafrost thawing or  
 291 flooding, which are in the form of methane (CH<sub>4</sub>). Considering the residence time of CO<sub>2</sub> and  
 292 methane in the atmosphere, the instantaneous value of the dynamic characterization factor (expressed  
 293 as time-integrated increase in CO<sub>2</sub>-equivalent concentration) of a pulse emission at any year  $T$  after  
 294 the emission was calculated as:

$$296 \quad DCF_{\text{inst},i}(T) = \int_{t_{\text{ini}}-1}^{t_{\text{ini}}} \Delta C_i \cdot IRF_i(t) \cdot \frac{RE_i}{RE_{\text{CO}_2}} dt \quad \text{for } T = 1, 2, 3, \dots \quad (S8)$$

297  
 298 where  $DCF_{\text{inst},i}(T)$  is the instantaneous dynamic characterization factor of gas  $i$  computed for  
 299 intervals of 1-year length (i.e.  $T$  = year 1, year 2, etc.), where the relative time interval  $T$  is defined



within  $T = [t_{\text{ini}}, t_{\text{end}}]$ , and where time  $t$  is relative and continuous. The  $\Delta C_i$  ( $\text{ppm}_i \cdot \text{kg}_i^{-1}$ ) is the change in atmospheric GHG concentration due to a unit emission of gas  $i$ ;  $IRF_i(t)$  is the impulse response function of gas  $i$  representing the decay of the gas after a pulse emission (defined as in eq. S2, for non- $\text{CO}_2$  gasses, and eq. S3, for  $\text{CO}_2$ );  $RE_i$  is the radiative efficiency per ppm of gas  $i$  ( $0.37 \text{ W} \cdot \text{m}^{-2} \cdot \text{ppm}^{-1}$  for methane<sup>71</sup>) and  $RE_{\text{CO}_2}$  is the radiative efficiency per ppm  $\text{CO}_2$  (see Table S3). Note that  $DCF_{\text{inst},i}$  is expressed as time-integrated increase in  $\text{CO}_2$ -equivalent concentration ( $\text{ppm CO}_2\text{e} \cdot \text{yr} \cdot \text{kg}_i^{-1}$ ) and thus it deviates from the  $DCF$  expressed as time-integrated radiative forcing increase ( $\text{W} \cdot \text{m}^{-2} \cdot \text{yr} \cdot \text{kg}_i^{-1}$ ) calculated in Levasseur et al.<sup>70</sup>. The  $\Delta C_i$  was calculated as:

$$\Delta C_i = \frac{1 \cdot 10^6 / M_i}{m_{\text{air}} / M_{\text{air}}} \cdot 1 \cdot 10^3 \quad (\text{S9})$$

Where  $M_i$  is the molar mass of gas  $i$  ( $16$  and  $44 \text{ g} \cdot \text{mol}^{-1}$  for  $\text{CH}_4$  and  $\text{CO}_2$  respectively);  $m_{\text{air}}$  ( $5.14 \cdot 10^{21} \text{ g}$ ) and  $M_{\text{air}}$  ( $28.9 \text{ g} \cdot \text{mol}^{-1}$ ) are the total mass and the molar mass of air in the atmosphere respectively<sup>71</sup>,  $1 \cdot 10^6$  (ppm) and  $1 \cdot 10^3$  ( $\text{g} \cdot \text{kg}^{-1}$ ) are conversion factors. The instantaneous dynamic characterization factor represents the annual value of the characterization factor of a pulse emission, where the pulse emission is the annual amount of  $\text{CO}_2$  or methane emissions caused by crossing a tipping point. By combining the instantaneous dynamic characterization factor of one annual release with the evolution of this annual release over the years, the dynamic effect at year  $T_{\text{effect}}$  was obtained:

$$C_{\text{tip},a}(T_{\text{effect}}) = \sum_i \sum_{T=T_a+1}^{T_{\text{effect}}} e_i(T) \cdot DCF_{\text{inst},i}(T_{\text{effect}} - T) \quad (\text{S10})$$

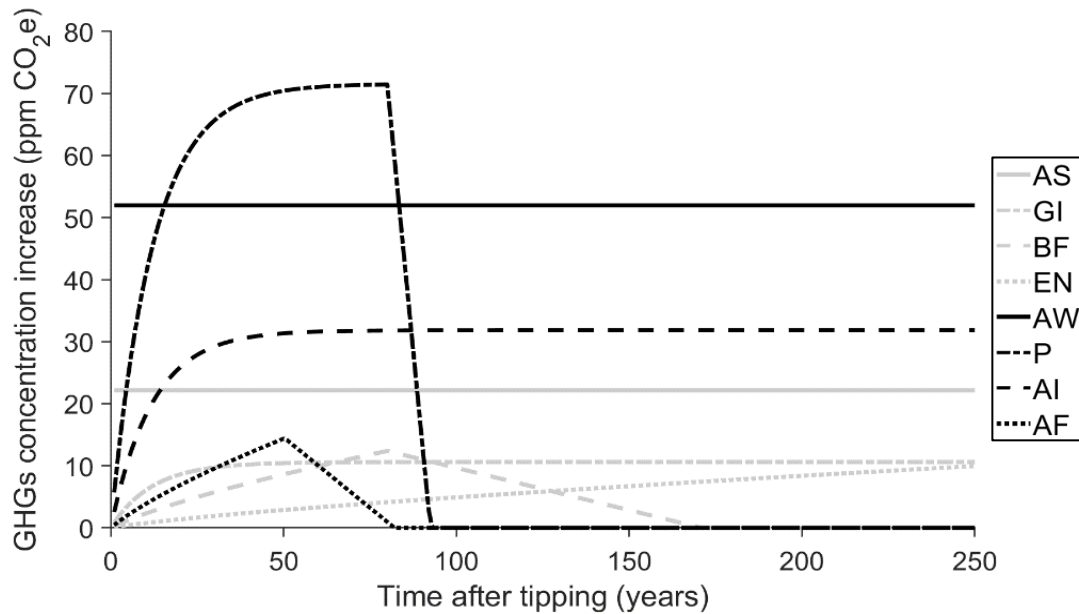
Where  $T_a$  is the year of tipping of the element  $a$ ,  $e_i(T)$  ( $\text{kg}_i \cdot \text{yr}^{-1}$ ) is the annual release of gas  $i$  (either  $\text{CO}_2$  or methane depending on the tipping point) at year  $T$ , i.e. at any year before the year of calculation of the dynamic effect ( $T_{\text{effect}}$ ) since the year following the tipping ( $T_a + 1$ ). The  $e_i(T)$  is determined by equally subdividing the estimated total carbon release after tipping (shown in Table S6) over the transition period of the tipping event (i.e. time required for the full effect to unfold found in Table S1). In this way, the time evolution of the annual release is assumed to be constant, meaning that the same amount of emissions is released every year over the transition period. In the case of GI, AI and Amazon forest, the transition period was also considered to calculate for how long this annual release lasts (e.g. for Amazon tipping point annual release ceases after 50 years). For EN the estimated release is basically permanent<sup>10</sup>, while for permafrost and Boreal forest 80 years of release

are assumed given that the figures used from ref. <sup>5</sup> refer to emissions from present day to 2100 ( $\approx 80$  years)<sup>5</sup>.

At any year after tipping ( $T_{\text{effect}}$ ), the dynamic effect calculated through eq. S10 is the result of the emissions released at year  $T_{\text{effect}}$  and the non-decayed fraction of emissions that occurred in all previous years ( $T$ ) since the year of tipping. The result provides the cumulative increase in CO<sub>2</sub>-equivalent concentration at time  $T_{\text{effect}}$  caused by every discrete annual release occurred since the year after the tipping event until  $T_{\text{effect}}$ . Dynamic and constant (for Arctic summer and winter sea ice loss) effects from tipping used in this study are presented in Figure S2.

**Table S6:** Best available estimated consequences from tipping found in the literature used to calculate  $C_{\text{tip}}$  for 8 tipping elements (no effect was modelled for the other tipping elements).

Selected tipping element	Estimated consequences from tipping		
	Amount	Unit	Definition
Arctic summer sea ice loss (AS)	0.29	$\text{W} \cdot \text{m}^{-2} \cdot \text{yr}^{-1}$	Total annual radiative forcing caused by one month of ice-free conditions, calculated from reduced sea-ice albedo effect <sup>72</sup> . In this study, the effect is assumed to unfold completely the year after tipping and to remain constant over time.
Greenland ice sheet melt (GI)	100	Gt C	Total carbon emissions, released over the transition period (best estimate 1500 years), from flooding of large areas of low-lying permafrost <sup>10</sup> .
West Antarctic ice sheet collapse (AI)	100	Gt C	Total carbon emissions, released over the transition period (best estimate 500 years), from flooding of large areas of low-lying permafrost <sup>10</sup> .
Amazon rainforest dieback (AF)	50	Gt C	Total carbon emissions, released over the transition period (best estimate 50 years), from forest dieback <sup>10</sup> .
Boreal forest dieback (BF)	10-40 (30)	Gt C	Estimated carbon emissions by 2100 <sup>5</sup> . In this study, it is assumed to be the total effect from tipping occurring over a transition period of 80 years.
El Niño-Southern Oscillation change in amplitude (EN)	0.2	$\text{Gt C} \cdot \text{yr}^{-1}$	Total annual carbon emissions released from anomalous fire events caused by stronger El Niño events. Assumed to be a permanent effect <sup>10</sup> .
Permafrost loss (P)	20-80 (45)	Gt C	Estimated carbon emissions by 2100 <sup>5</sup> . In this study, it is assumed to be the total effect from tipping occurring over transition period of 80 years.
Arctic winter sea ice loss (AW)	0.68	$\text{W} \cdot \text{m}^{-2} \cdot \text{yr}^{-1}$	Total annual radiative forcing caused by year-round ice-free conditions, calculated from reduced sea-ice albedo effect <sup>72</sup> . This effect is assumed to unfold completely the year after tipping and to remain constant over time.



**Figure S2:** Temporal evolution of CO<sub>2</sub>-equivalent concentration increase caused by crossing eight of the selected tipping elements as modelled in this study. Uncertainty surrounding these results is undefined and thus not considered. AS = Arctic summer sea ice loss, GI = Greenland ice sheet melt, BF = Boreal forest dieback, EN = El Niño-Southern Oscillation change in amplitude, AW = Arctic winter sea ice loss, P = Permafrost loss, AI = West Antarctic ice sheet collapse, AF = Amazon rainforest dieback.

### S1.3.2 Notes on tipping elements for which $C_{tip}$ was not modelled

**Sahara/Sahel and West African monsoon shift (AM):** It is uncertain whether AM will shift northward (leading to increased rainfall) or southward (leading to further drying of the Sahel)<sup>3</sup>. In either cases, modelling the consequences of a local change in rainfall on the global climate system remains challenging and no estimate allowing expressing the effect as a change in CO<sub>2</sub>-equivalent concentration was found. Therefore, the effect from crossing this climate tipping point was not modelled.

**Atlantic thermohaline circulation shutoff (TC) and North Atlantic Subpolar Gyre convection collapse (SG):** The main expected consequences of a potential Atlantic thermohaline circulation shutoff are cooling of the northern hemisphere and warming of the southern hemisphere<sup>21,73,74</sup>. A collapse of the subpolar gyre convection would cause local cooling over the North Atlantic and, to a minor extent, over Western Europe and North American coast<sup>21</sup>. Some studies found that a collapse of both TC and SG would cause a reduction of the global mean temperature, leading to an overall cooling of the planet contrasting the global warming trend<sup>21,75</sup>. However, no clear evidence of the underlying mechanism emerges from the studies and the same authors highlight that the observed cooling might be the result of factors such as the climate sensitivity parameter used in the models<sup>21</sup>. For this reason, no effect from these two tipping points is assumed in our model.

**Alpine glaciers loss (AG):** Shrinkage of Alpine glaciers and snow cover is expected to have mainly local effects, reducing surface reflectivity and thus leading to amplified temperature increase in the region<sup>26</sup>. However, no estimate allowing expressing the effect as a change in global CO<sub>2</sub>-equivalent concentration was found; therefore the effect from crossing this climate tipping point was not modelled.

**Coral reefs deterioration (CR):** Deterioration of coral reefs damages local marine habitats and species, causing loss of biodiversity<sup>27</sup>, however it does not have direct or measurable consequences on the climate system, such as temperature increase due to positive feedbacks. Thus, the effect from crossing this climate tipping point was not accounted for.

### ***S1.3.3 Atmospheric capacity cutoff***

To define a meaningful minimum value for the atmospheric capacity (i.e. the value below which the difference between CO<sub>2</sub>-equivalent concentrations is considered too small), the uncertainty surrounding the calculation of the capacity shall be quantified. The annual variability of atmospheric CO<sub>2</sub> concentrations, based on average in situ air measurements between 1959 – 2018 from Mauna Loa Observatory, Hawaii<sup>68</sup>, was used as a proxy. The average difference in annual CO<sub>2</sub> concentrations was found to be 5.89 ppm CO<sub>2</sub>, which was rounded to 6 ppm CO<sub>2</sub> because differences in annual measurements were not detected at decimal level. The cutoff was then applied to the difference between the concentration at the year of tipping and the concentration at year  $T$  when calculating the capacity as in eq. 4 in the main article.

Ideally, the uncertainty of future projections of GHGs concentrations should be used, as the capacity depends mainly on this parameter. However, when calculated from IPCC's temperature projections uncertainties<sup>76</sup>, we found cutoff values between 43 – 70 ppm CO<sub>2</sub> equivalents, depending on the RCP pathway, which are deemed too high and were therefore not used. As a measure of the variability of GHGs concentration within the yearly time step used in the model, CO<sub>2</sub> was taken as a proxy because it is the most abundant anthropogenic greenhouse gas in the atmosphere.

## S1.4 Consideration of time as a variable

**Table S7:** Summary of the symbols used to indicate the time variable in this paper.

Symbol	Meaning	Where it is used
$T$	<i>Interval-like</i> time variable indicating generic years that can be either <i>absolute</i> or <i>relative</i> .	In eq. 2, 3, 4, S1, S6, S8, S10.
$t$	<i>Point-like</i> time variable indicating continuous time that can be either <i>absolute</i> or <i>relative</i> .	In the standard definition of RF (eq. S2, S3) and as integration variable in eq. S8. Used also in Figure 1 as <i>absolute</i> variable.
$T_{\text{emission}}$	<i>Interval-like</i> and <i>absolute</i> time variable used to indicate <i>emission</i> years.	In all situations where the dependent variable is a function of the emission year and where years are intended as time intervals (eq. 1-4, S1 and S4-S6).
$T_{\text{tipping},j}$	<i>Interval-like</i> and <i>absolute</i> time variable indicating the year of tipping of the $j$ th expected tipping point.	In eq. 3, 4, S4-S6.
$T_{\text{effect}}$	<i>Interval-like</i> and <i>absolute</i> time variable used to indicate the years after crossing tipping points, when <i>effects</i> of tipping unfold.	For calculation of the concentration increase from tipping $C_{\text{tip}}$ (eq. S7 and S10).
$T_{\text{tipping},j_{\text{last}}}$	<i>Interval-like</i> and <i>absolute</i> time variable indicating the year of tipping of the last ( $j_{\text{last}}$ ) expected tipping point across 10,000 Monte Carlo simulations.	As upper bound of the summation sign in eq. 5.
$T_a$	<i>Interval-like</i> and <i>absolute</i> time variable indicating the year of tipping of tipping element $a$ (used for calculation of the effect from tipping). Subscript $a$ is used rather than subscript $j$ (as in $T_{\text{tipping},j}$ ), because the effect from tipping depends on the tipping element being crossed and not on its order of occurrence.	Used to define the lower bound of the summation in eq. S10

## S1.5 Example calculation of MCTP with two tipping points

Figure 1 in the main article illustrates how the MCTP framework can be applied with two tipping points. When considering the first tipping point ( $j = 1$ ) occurring at year  $T_{\text{tipping},1}$ , every emission at  $T_{\text{emission}} < T_{\text{tipping},1}$  (before  $T_{\text{tipping},1}$ ) takes up a certain part of the atmospheric capacity available before reaching the concentration at the year of tipping  $T_{\text{tipping},1}$  ( $C(T_{\text{tipping},1})$ ). However, when considering a second tipping point ( $j = 2$ ) at year  $T_{\text{tipping},2}$ , the same emission  $T_{\text{emission}} < T_{\text{tipping},1}$  will now take up also the capacity left before reaching the concentration at the second year of tipping  $T_{\text{tipping},2}$  ( $C(T_{\text{tipping},2})$ ). Therefore, the total MCTP of gas  $i$  emitted at year  $T_{\text{emission}} < T_{\text{tipping},1}$  is given by the sum of the MCTPs for the first and the second tipping point (first and second term in eq. S11 respectively). Assuming, hypothetically,  $T_{\text{tipping},1} = 2030$ ,  $T_{\text{tipping},2} = 2043$  and  $T_{\text{emission}} < T_{\text{tipping},1} = 2025$ :

$$\begin{aligned}
418 \quad MCTP_i(2025) &= \frac{I_{\text{emission},i,1}(2025)}{CAP_1(2025)} + \frac{I_{\text{emission},i,2}(2025)}{CAP_2(2025)} = \frac{\left\{ \frac{A_i \tau_i \left[ 1 - e^{-\frac{(2030-2025)}{\tau_i}} \right]}{RE_{CO_2}} \right\}}{C(2030) \cdot (2030 - 2025) - \sum_{k=1}^5 [C(T_{k-1})] \cdot 1} \\
419 \quad &+ \frac{\left\{ \frac{A_i \tau_i \left[ 1 - e^{-\frac{(2043-2025)}{\tau_i}} \right]}{RE_{CO_2}} \right\}}{(2043) \cdot (2043 - 2025) - \sum_{k=1}^{18} [C(T_{k-1})] \cdot 1} \quad (S11)
\end{aligned}$$

420

421 where, each term of the sum is the ratio between the *impact of the emission* for either the first or the  
422 second tipping point and the corresponding remaining capacity;  $A_i$  is the radiative efficiency of gas  $i$   
423 [ $\text{W} \cdot \text{m}^{-2} \cdot \text{kg}_i^{-1}$ ];  $\tau_i$  is the atmospheric lifetime of gas  $i$  [yr] and  $RE_{CO_2}$  is the radiative efficiency per  
424 ppm  $CO_2$  [ $\text{W} \cdot \text{m}^{-2} \cdot \text{ppm } CO_2^{-1}$ ]. Note, that the  $C_{\text{tip}}(T)$  term is not included in eq. S11 because no  
425 tipping point has been crossed yet in this case.

426 For an emission at  $T_{\text{emission}} > T_{\text{tipping},1}$  (after  $T_{\text{tipping},1}$ ), which will also take up part  
427 of the capacity left before  $T_{\text{tipping},2}$ , but has no influence on  $T_{\text{tipping},1}$ , the total MCTP is given only  
428 by the contribution to exceed the second tipping point (assuming  $T_{\text{emission}} > T_{\text{tipping},1} = 2035$ ):

429

$$\begin{aligned}
430 \quad MCTP_i(2035) &= \frac{I_{\text{emission},i,2}(2035)}{CAP_2(2035)} \\
431 \quad &= \frac{\left\{ \frac{A_i \tau_i \left[ 1 - e^{-\frac{(2040-2035)}{\tau_i}} \right]}{RE_{CO_2}} \right\}}{(2040) \cdot (2040 - 2035) - \sum_{k=1}^5 [C(T_{k-1}) + C_{\text{tip}}(T_{k-1})] \cdot 1} \quad (S12)
\end{aligned}$$

432

433 Here, the tipping point occurring at  $T_{\text{tipping},1}$  has an effect on the climate system ( $C_{\text{tip}}$ ) that further  
434 reduces the remaining capacity up to the following year of tipping and anticipating tipping from 2043  
435 to 2040 (so that  $T_{\text{tipping},2} = 2040$ ). This is also accounted for in the calculation of the capacity.

436 While for both emissions at  $T_{\text{emission}} < T_{\text{tipping},1}$  and  $T_{\text{emission}} > T_{\text{tipping},1}$  the *impact of the*  
437 *emission* depends on the residence time of the gas in the atmosphere and the difference between the  
438 emission year and the year of tipping, the capacity varies depending on i) the difference in  
439 atmospheric  $CO_2$ -equivalent concentration between the emission year and the year of tipping and ii)  
440 on timing of emissions. For emissions at  $T_{\text{emission}} < T_{\text{tipping},1}$  (eq. S11), the effect from tipping ( $C_{\text{tip}}$ )  
441 is equal to zero as no tipping points have been crossed yet and therefore the total remaining capacity  
442 for these emissions is not affected. On the contrary, for emissions at  $T_{\text{emission}} > T_{\text{tipping},1}$  (eq. S12),  
443 the remaining capacity is influenced by the  $C_{\text{tip}}$  from the crossed level.

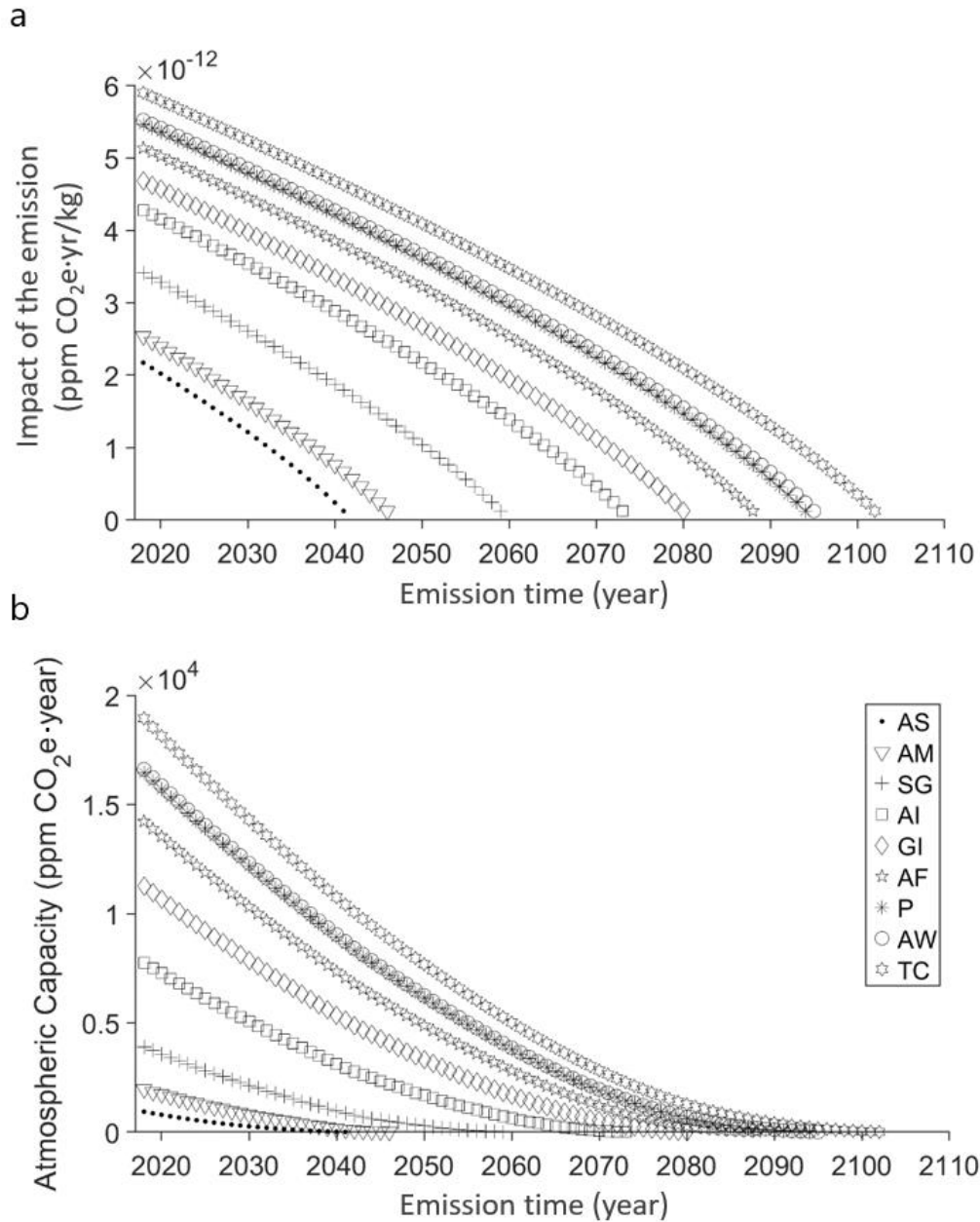
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## S1.6 Details of two illustrative simulations

This section gives an overview on occurrence of tipping points in the two illustrative simulations presented in the main article and shows typical trends for *impacts of the emission* and remaining capacities in the calculation of MCTPs (taking as example simulation 1).

**Table S8:** Data on tipping points triggered in simulation 1 and simulation 2.

	Triggered tipping points (from first to last)	Threshold temperature (°C above pre-industrial)	Concentration at year of tipping (ppm CO <sub>2e</sub> )	Anticipated year of tipping (year)	Year of tipping without $C_{tip}$ (year)
Simulation 1	Arctic summer sea ice loss	1.62	484	2042	2042
	West African monsoon shift	1.87	523	2047	2054
	North Atlantic subpolar gyre convection collapse	2.17	576	2060	2065
	West Antarctic ice sheet collapse	2.59	652	2074	2078
	Greenland ice sheet melt	2.95	711	2081	2092
	Amazon rainforest dieback	3.23	755	2089	2107
	Permafrost loss	3.41	784	2095	2119
	Arctic winter sea ice loss	3.42	786	2096	2120
	Atlantic thermohaline circulation shutoff	3.75	814	2103	2165
Simulation 2	Arctic summer sea ice loss	1.95	535	2056	2056
	West African monsoon shift	2.17	576	2060	2065
	Greenland ice sheet melt	2.30	600	2065	2069
	North Atlantic subpolar gyre convection collapse	2.40	617	2067	2072
	Arctic winter sea ice loss	2.59	652	2073	2078
	Atlantic thermohaline circulation shutoff	3.34	773	2100	2114
	Permafrost loss	4.02	814	2116	2298



**Figure S3:** (a) *Impact of the emission* of 1 kg CO<sub>2</sub> and (b) remaining atmospheric capacity relative to each of the 9 tipping points triggered in simulation 1. Both *impacts* and remaining capacities decrease over time, but capacities decrease faster than the *impacts*, explaining why MCTPs increase while approaching a tipping point. AS = Arctic summer sea ice loss, AM = West African monsoon shift, SG = North Atlantic subpolar gyre convection collapse, AI = West Antarctic ice sheet collapse, GI = Greenland ice sheet melt, AF = Amazon rainforest dieback, P = Permafrost loss, AW = Arctic winter sea ice loss, TC = Atlantic thermohaline circulation shutoff.



## S1.7 Calculation of impact scores for the case study

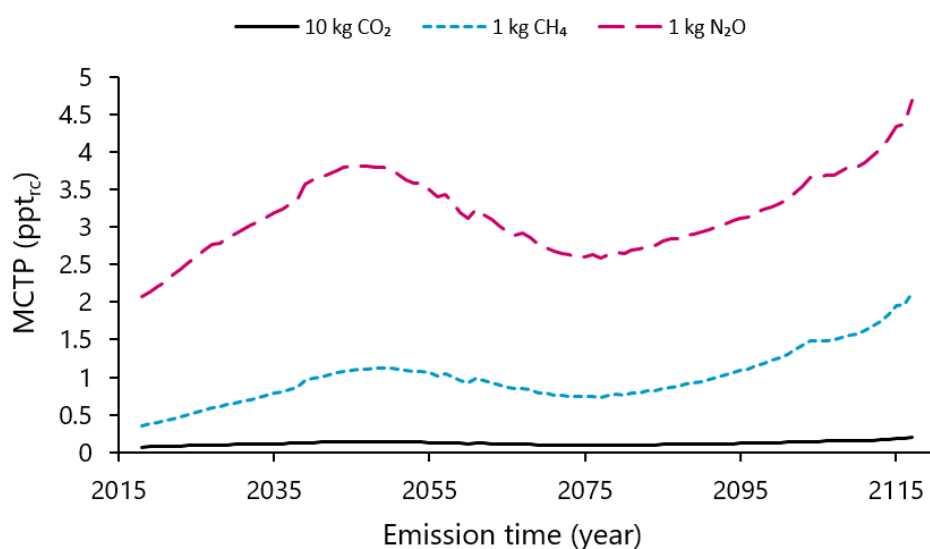
**Table S9:** Overview of equations used for calculation of impact scores (*IS*) with all metrics used in this study. For all metrics, *i* denotes a specific GHG and  $T_{\text{emission}}$  the emission year. Note that the original notation for the time variable of the GWP-based and GTP CFs was harmonized with the notation used in this paper.

	Impact score calculation	Symbols
MCTP [ppt <sub>rc</sub> /kg plastic]	$IS_{\text{MCTP}} = \sum_i \sum_{T_{\text{emission}}=2021}^{T_{\text{tipping},j_{\text{last}}}} m_i(T_{\text{emission}}) \cdot MCTP_i(T_{\text{emission}})$	$T_{\text{tipping},j_{\text{last}}}$ = year of last tipping point; $m_i(T_{\text{emission}})$ = mass of gas <i>i</i> emitted at year $T_{\text{emission}}$ [kg]; $MCTP_i(T_{\text{emission}})$ = MCTP for gas <i>i</i> and emission year $T_{\text{emission}}$ [ppt <sub>rc</sub> · kg <sup>-1</sup> ]
GWP20 [kg CO <sub>2</sub> eq/kg plastic]	$IS_{\text{GWP20}} = \sum_i M_i \cdot GWP20_i$	$M_i$ = total mass of gas <i>i</i> emitted in 20 years [kg]; $GWP20_i$ = GWP20 of gas <i>i</i> [kg CO <sub>2</sub> eq · kg <sup>-1</sup> ]
GWP100 [kg CO <sub>2</sub> eq/kg plastic]	$IS_{\text{GWP100}} = \sum_i M_i \cdot GWP100_i$	$M_i$ = total mass of gas <i>i</i> emitted in 100 years [kg]; $GWP100_i$ = GWP100 of gas <i>i</i> [kg CO <sub>2</sub> eq · kg <sup>-1</sup> ]
GWP100 <sub>ILCD</sub> [kg CO <sub>2</sub> eq/kg plastic]	$IS_{\text{GWP100ILCD}} = \sum_i M_i \cdot GWP100_i - Credit_i$ $Credit_i = \sum_{T=2}^{100} m_i(T) \cdot T \cdot EQ_i$	$Credit_i$ = ILCD credit for carbon storage; $m_i(T)$ = mass of gas <i>i</i> emitted at relative time <i>T</i> ; $EQ_i$ = equivalency factor for gas <i>i</i> (0.01, 0.34 and 0.36 [kg CO <sub>2</sub> eq · kg <sup>-1</sup> · yr <sup>-1</sup> ] for CO <sub>2</sub> , biogenic and fossil CH <sub>4</sub> respectively <sup>77,78</sup> )
Dynamic GWP100 [kg CO <sub>2</sub> eq/kg plastic]	$IS_{\text{dynGWP100}} = \frac{GWI_{\text{cum}}(100)}{\int_0^{100} A_{\text{CO}_2} \cdot C(t)_{\text{CO}_2} dt}$ $GWI_{\text{cum}}(100) = \sum_{T=0}^{100} GWI_{\text{inst}}(T)$ $GWI_{\text{inst}}(100) = \sum_i \sum_{T=0}^{100} m_i(T) \cdot DFC_i(100 - T)$ $DFC_i(100) = \int_{t-1}^{100} A_i \cdot C(t)_i dt$	$GWI_{\text{cum}}(100)$ = cumulative global warming impact in 100 years [ $\text{W} \cdot \text{m}^{-2} \cdot \text{yr} \cdot \text{kg}_i^{-1}$ ]; $A_{i/\text{CO}_2}$ = radiative efficiency per unit mass gas <i>i</i> /CO <sub>2</sub> increase [ $\text{W} \cdot \text{m}^{-2} \cdot \text{kg}_i^{-1}$ ]; <i>T</i> = relative <i>interval</i> time [yr]; <i>t</i> = relative <i>point in time</i> [yr]; $C(t)_{i/\text{CO}_2}$ = atmospheric load of gas <i>i</i> /CO <sub>2</sub> at time <i>t</i> after emission [kg]; $GWI_{\text{inst}}(100)$ = instantaneous global warming impact in 100 years [ $\text{W} \cdot \text{m}^{-2} \cdot \text{yr} \cdot \text{kg}_i^{-1}$ ]; $DFC_i$ = dynamic characterization factor of gas <i>i</i> [ $\text{W} \cdot \text{m}^{-2} \cdot \text{yr}$ ]; (see equations 1-4 in Levasseur et al. <sup>79</sup> )
GTP100 [kg CO <sub>2</sub> eq/kg plastic]	$IS_{\text{GTP100}} = \sum_i M_i \cdot GTP100_i$	$M_i$ = total mass of gas <i>i</i> emitted over 100 years [kg]; $GTP100_i$ = GTP100 of gas <i>i</i> [kg CO <sub>2</sub> eq · kg <sup>-1</sup> ]

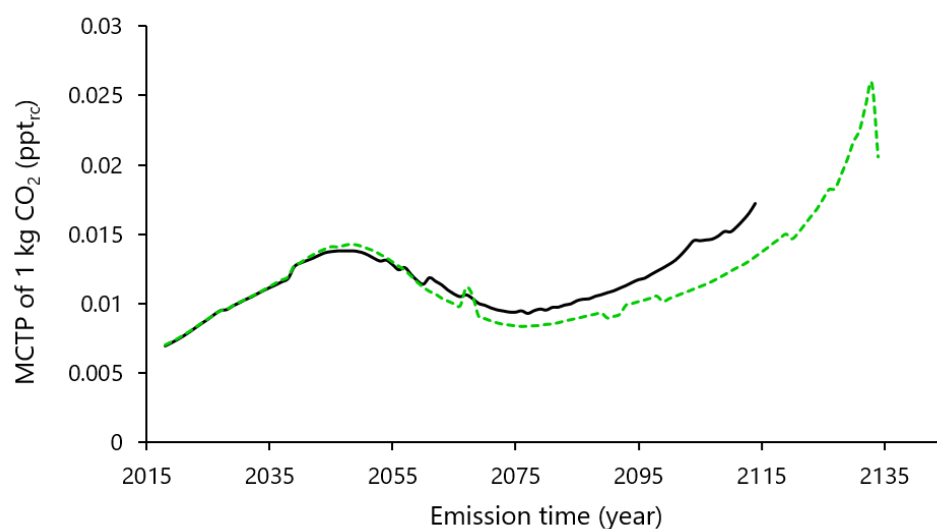
## S2. Supplementary results

Sub-section S2.1 contains additional figures comparing average MCTP results for the three major anthropogenic GHGs (carbon dioxide (CO<sub>2</sub>), methane (CH<sub>4</sub>) and nitrous oxide (N<sub>2</sub>O)) and without considering the effects from tipping. In sub-section S2.2, we show details on the case study, comparing emission profiles with MCTP values and presenting MCTP impact scores calculated under different RCP pathways.

### S2.1 Supplementary MCTP results

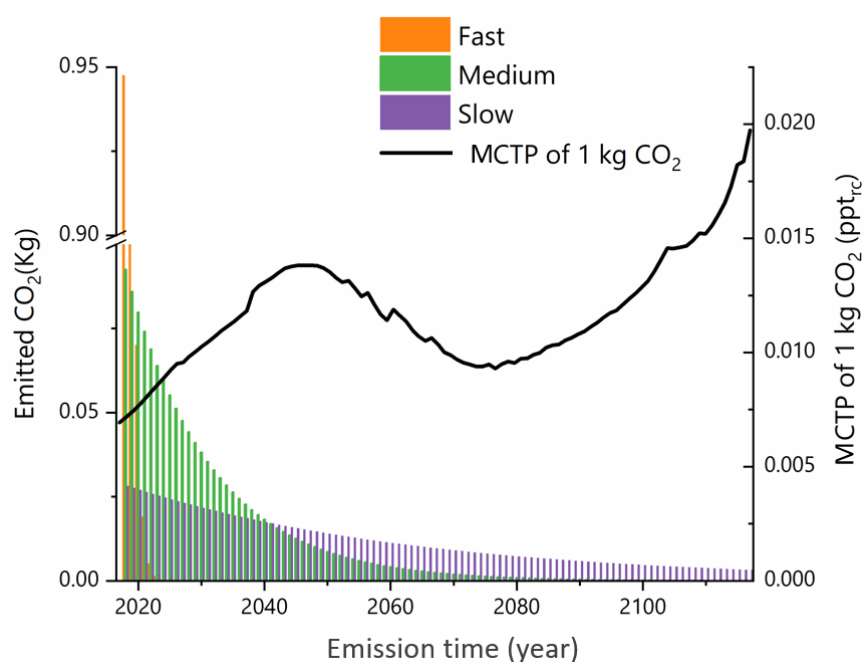


**Figure S4:** Average (geometric mean) multiple climate tipping points potential (MCTP) of CO<sub>2</sub>, CH<sub>4</sub> and N<sub>2</sub>O at different emission years under RCP6 pathway. Note that results for CO<sub>2</sub> are given for 10 kg rather than 1 kg.



**Figure S5:** Average (geometric mean) multiple climate tipping points potential (MCTP) of 1 kg CO<sub>2</sub> under RCP6 considering effects from crossing tipping points (black solid line) and without considering such effects (green dashed line).

## S2.2 Details and supplementary results for the case study



**Figure S6:** CO<sub>2</sub> emission profile for fast, medium and slow degrading plastics under anaerobic conditions (left axis) and multiple climate tipping points potential (MCTP) per unit CO<sub>2</sub> emission (right axis).

**Table S10:** Climate tipping impact scores for different end-of-life degradation scenarios of plastic polymers calculated assuming background concentration pathways RCP4.5, RCP6 and RCP8.5. Ranking between scenarios 1 – 7 is illustrated within each column with different colors. Red shading indicates the highest impact scores and green the lowest impact scores.

End-of-life degradation scenario	MCTP - under RCP4.5 (ppt <sub>re</sub> /kg plastic)	MCTP - under RCP6 (ppt <sub>re</sub> /kg plastic)	MCTP - under RCP8.5 (ppt <sub>re</sub> /kg plastic)
1. Incineration <sup>a</sup>	0.015	0.014	0.014
Plastic degradation rate <sup>b</sup>			
2. Fast <sup>c</sup>	0.12	0.089	0.11
3. Medium <sup>d</sup>	0.22	0.14	0.17
4. Slow <sup>e</sup>	0.25	0.16	0.13
5. Very slow <sup>f</sup>	0.0027	0.0020	0.0011
Delayed degradation <sup>g</sup>			
6. After 20 years (fast rate)	0.31	0.21	0.24
7. After 50 years (fast rate)	0.45	0.16	0.15

<sup>a</sup>Incineration of fossil-based plastic where all carbon is emitted as CO<sub>2</sub> in the first year. <sup>b</sup>Degradation under anaerobic conditions. <sup>c</sup>90% degradation of polycaprolactone (PCL) in 2 years<sup>80</sup>. <sup>d</sup>90% degradation of polybutylene succinate (PBS) in 31 years<sup>81</sup>. <sup>e</sup>90% degradation of polystyrene (PS) in 105 years<sup>82</sup>. <sup>f</sup>1% degradation of bio-based PLA in 100 years<sup>83</sup>. <sup>g</sup>Potential short (20 years) and long (50 years) lag phase in degradation based on ref. <sup>83</sup>.

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