Supporting Information:

Holocene variability and Anthropocene rates of change

Understanding the Anthropocene requires that it be analyzed in the context of the dynamics of the Earth System, the complex system consisting of a highly intertwined geosphere and biosphere (1). Here we focus on the late Quaternary, the past 1.2 million years, encompassing the period in which modern humans evolved and developed (1). In particular, we focus on the Holocene, the most recent interglacial period that has proven to be especially accommodating for the development of human society.

The climate of the late Quaternary is characterized by quasi-periodic ‘saw-tooth’ oscillations between two reasonably well-defined bounding states - full glacial and interglacial states defined by multi-millennial extremes in global mean temperature, sea level, continental ice volume and atmospheric composition. The Holocene, the most recent interglacial that began approximately 11,700 years before present, was relatively stable compared to the preceding transition from the Last Glacial Maximum, a transition accompanied by a sharp rise in temperature and a ~130 m rise in sea level. During the Holocene, *Homo sapiens* created complex societies that modify and exploit their environments with ever-increasing scope, from hunter-gatherer and agricultural communities to the highly technological, rapidly urbanizing societies of the 21st century.

However, there was considerable variability through the Holocene, particularly at the regional level, which often posed significant challenges for human societies (Section “What Is at Stake?” in the main text). Holocene variability, especially pronounced in the Northern Hemisphere, was driven primarily by changes in orbital insolation, solar output, and albedo (especially reflection of solar radiation by ice sheets). Orbital insolation reached its peak in the Northern Hemisphere around 11,000 years ago, but much of this incoming solar radiation went into melting the remains of the great Northern Hemisphere ice sheets. The Holocene thermal optimum was not reached until the ice sheets (except Greenland) had gone, around 7,000 years ago, by which time orbital insolation was well into its decline towards current levels. Sea levels were low but rising, reaching their modern extent around 7,000 years ago (2). Warm conditions south of the ice sheets, in places like the Middle East, accompanied by rains from a more extended Inter-Tropical Convergence Zone (ITCZ), supported the development of agriculture there and hunting and pastoralism in what is now the Sahara Desert. With the gradual cooling caused by declining orbital insolation, the ITCZ shrank back, removing rains from the Sahara by 4,500 years ago, and weakening the monsoons. By 4,000 years ago the declining orbital insolation of the Northern Hemisphere was leading it into a ‘neoglacial’ characterized by glacial advance in mountain regions. According to Ruddiman et al. (3), the expansion of agriculture from about 6,000 years ago may have emitted enough CO_2 from deforestation to have prevented the Holocene neoglacial from being colder than it would have been otherwise.

That decline in orbital insolation gradually flattened, is now almost imperceptible, and is projected to stay flat for the next 1,000 years (4). Superimposed on the decline were small changes in solar output (5), which led to the Roman Warm Period and the Medieval Warm Period. Because the underlying orbital insolation baseline was in decline, the Roman Warm Period was somewhat warmer than the Medieval Warm Period. Intervening weak periods in solar output (5) led to cooling, commonly associated with rains across western Europe (6,7). In recent times the strongest of these cool periods was the Little Ice Age (1250-1850 CE),

www.pnas.org/cgi/doi/10.1073/pnas.1810141115

which comprised four main cold periods separated by times that were nearly as warm as the Medieval Warm Period. Occasional volcanic eruptions large enough to eject dust into the stratosphere led to brief periods of cooling like that of Tambora in 1815, which led to the 'year without a summer' of 1816. Regional changes in hydroclimate, many of which were much more pronounced than changes in global average conditions (e.g., 6,7), were also a frequent feature of the Holocene, often with serious consequences for the wellbeing of human societies at the time (Section "What Is at Stake?" in the main text).

The Anthropocene represents the beginning of a very rapid human-driven trajectory of the Earth System away from a globally stable Holocene state towards new, hotter climatic conditions and a profoundly different biosphere (8,9,10). The current atmospheric CO_2 concentration of over 400 ppm is already well above the Holocene maximum and, indeed, the upper limit of any interglacial (11). The global mean temperature rise based on a 30-year average is ~0.9°C above pre-industrial (derived from 12), and the global mean temperature anomaly for the 2015-2017 period is over 1^{o} C above a pre-industrial baseline (13), likely close to the upper limit of the past interglacial envelope of variability.

Although rates of change in the Anthropocene are necessarily assessed over much shorter periods than those used to calculate long-term baseline rates and therefore present challenges for direct comparison, they are nevertheless striking. The rise in global CO_2 concentration since 2000 is about 20 ppm/decade, which is up to 10 times faster than any sustained rise in CO_2 during the past 800,000 years (14). Since 1970 the global average temperature has been rising at a rate of 1.7°C per century, compared to a long-term decline over the past 7,000 years at a baseline rate of 0.01°C per century (12,15).

These current rates of human-driven changes far exceed the rates of change driven by geophysical or biosphere forces that have altered the Earth System trajectory in the past (e.g., 16,17); even abrupt geophysical events do not approach current rates of human-driven change. For example, the Paleocene-Eocene Thermal Maximum (PETM) at 56 Ma BP (before present), a warming that reached 5-6°C and lasted about 100,000 years, accompanied by a rise in sea level and ocean acidification, drove the extinction of 35-50% of the deep marine benthic foraminifera and led to continent-scale changes in the distributions of terrestrial plants and animals (18-21). The warming was driven by a carbon release estimated to be ~1.1 Gt C y^{-1} (14). While initial estimates suggested a cumulative carbon emission of 2,500-4,500 Gt C at the PETM (22), a more recent analysis suggests that the emission was closer to 10,000 Gt C, much of it coming from eruptions in the North Atlantic Igneous Province (23). By comparison, the current human release of carbon to the atmosphere is nearly an order-of-magnitude greater at ~10 Gt C y^{-1}(24). Cumulative human emissions of CO_2 from 1870 through 2017 have reached ~610 Gt C (24).

The fastest rates of past change to the biosphere were caused by rare catastrophic events, for instance, a bolide strike at 66 Ma BP that ended the Mesozoic Era. That event, which led to the demise of the dinosaurs and the rise of the age of mammals, was concurrent with the eruption of the Deccan Traps, one of a series of outpourings of lava, scattered through time, that formed vast expanses of plateau basalts in Large Igneous Provinces over periods perhaps as short as a hundred thousand years. Large Igneous Provinces are commonly associated with major periods of biological extinction (25). For example, around 252 million years ago at the end of the Permian Period the eruption of the Siberian Trap lavas is associated with the disappearance of up to 96% of marine species and 70% of land species, and recovery took millions of years (26).

In contrast, human activity has significantly altered the climate in less than a century (27), driving many changes within marine, freshwater and terrestrial ecosystems on multiple processes at different levels of biological organization (28). Species movements in response to climate change already rival or exceed rates of change at the beginning or the end of the Pleistocene, and may increase by an order of magnitude in the near future (29). In terms of their influence on the carbon cycle and climate, the human-driven changes of the Anthropocene are beginning to match or exceed the rates of change that drove past, relatively sudden mass extinction events (30), and are essentially irreversible (31).

Direct human-driven changes to the Anthropocene biosphere are probably even more profound than those in the climate system (e.g., 32). Humans have transformed 51% of Earth's land cover from forest and grassland to anthropogenic biomes of crops, cities and grazing lands (33), mostly since 1700 (34). Current extinction rates are at least tens and probably hundreds of times greater than background rates (35), and are accompanied by distributional and phenological changes that disrupt entire ecosystems both on land and in the sea. The rate of homogenization of biota (species translocations across the planet) has risen sharply since the mid-20th century (32,36).

In summary, the Anthropocene is in a strongly transient phase in which human societies, the biosphere and the climate system are all changing at very rapid rates.

Insights from time periods in Earth's past

Consideration of biogeophysical feedbacks and tipping cascades (Sections "Biogeophysical Feedbacks" and "Tipping Cascades" in the main text) raises the possibility that a threshold could be crossed that irrevocably takes the Earth System on a trajectory to a hotter state. To gain some orientation as to what that state could (or could not) be like, we consider four time periods in the Earth's recent past that resemble points along potential pathways over the next millennia (Figure 1, Table S1). Although these past states are not direct analogues for the future, given differences among them in astronomical forcing, topography, bathymetry, oceanic and atmospheric circulation, and biome characteristics, they may provide some insights into general features of the Earth System that could potentially occur at various points along future pathways. Global temperature differences across the interglacial periods of the late Quaternary (the first two comparative time periods of Table S1) are explained in part by the small changes in CO_2 concentration, and in part by seasonal forcing due to changes in Earth's orbit and obliquity. Forced and unforced natural modes of variability that operate at time scales of decades to millennia may also have contributed to the precise positioning of the pre-industrial baseline. Nevertheless, several robust insights emerge from this comparison.

First, conditions of the mid-Holocene and the Eemian are already inaccessible because CO_2 concentration is currently ~400 ppm and rising, and temperature is rising rapidly towards or past the upper bounds of these time periods. That is, the Earth System has likely departed from the glacial-interglacial limit cycle of the late Quaternary (42,43). This insight is consistent with an analysis of the Anthropocene from both Earth System science and stratigraphic perspectives (e.g., 31,44).

Second, given that the Earth is nearly as warm as Eemian conditions and current atmospheric CO_2 concentration has already reached the lower bound of mid-Pliocene levels (Table S1), an eventual sea-level rise of up to 10 m or more is likely unless CO_2 emissions are rapidly reduced and widespread CO_2 removal from the atmosphere is deployed.

Table S1: Level of warming and long-term sea-level rise for the four time periods marked on Figure 1.

Climate state/condition	Time before present	Atmos. CO_2 conc.	Global mean surface temp (°C)[1]	Sea-level rise in paleo-record (m)[2]	Likelihood of stabilizing near these conditions in the Anthropocene	References
Current (2017)	0	400	>1.0	Not applicable	Stability in current conditions not possible; further temperature, CO_2 and sea level rises locked in	13,37
A. Mid-Holocene	~6-7 ka	260	0.6-0.9	Not applicable	State not accessible; system moving away from these conditions	15
B. Eemian	~125 ka	280-300	1.0-1.5	6-9	State not accessible: CO_2 now much higher and still rising; temperature likely to stabilize at higher than Eemian level, sea-level rise may well be this high or higher	38
C. Mid-Pliocene	~3-4 Ma	up to 400-450	2-3	10-22	Possibly accessible but only if Paris 2°C target is met (a best-case scenario)	39
D. Mid-Miocene	~15-17 Ma	up to 300-500	4-5	10-60	Likely with high emissions scenario;	40,41

					current trajectory	

[1]Compared to a pre-industrial baseline ca. 200 years BP, not to a long-term Holocene baseline. Furthermore, during the Quaternary, part of the global temperature signal is explained by seasonal forcing due to changes in Earth's orbit and obliquity (47). The current temperature is the anomaly for 2015/2016. The current temperature based on a 30-year average is ~0.9°C above pre-industrial. Temperature estimates for the mid-Holocene are centennial averages, while those for the other intervals are millennial averages.
[2]Estimates for sea-level rise in the past are millennial, or longer, averages. There is considerable uncertainty around paleo-estimates of sea-level rise (compared to Holocene), primarily due to uncertainties in dating and non-eustatic processes, such as local uplift and isostasy.

Finally, Table S1 emphasizes that we may be fast approaching a commitment to Pliocene-like conditions, which lie beyond the threshold in Figure 1. If the Earth System moves onto a 'Hothouse Earth' pathway towards conditions resembling the mid-Miocene, maximum greenhouse conditions would be reached in several centuries (2). After peak warming, transient changes in alkalinity balance following the dissolution of deep-sea carbonates (on about a 10,000-year timescale) and silicate weathering (on about a 100,000-year timescale) would eventually reduce the global mean surface temperature back to pre-industrial levels, assuming no significant changes in other forcing during that period. However, it would take on the order of 100,000 years for conditions to return to their pre-perturbation levels based on the trajectory followed after the Paleocene-Eocene Thermal Maximum (45,46; broken red line in Figure 1).

Estimation of biogeophysical feedback strength (Section "Biogeophysical Feedbacks" in the main text)

Table S2: Biogeophysical feedbacks in the Earth System that could accelerate the trajectory towards a 'Hothouse Earth' pathway. The table includes feedbacks to both temperature and sea-level, the axes of Figure 1.

Feedback	Level of forcing considered for estimation of feedback strength	Strength of feedback[1]	Speed of Earth System response	Notes and references; see S1 for more details on estimation of feedbacks
Permafrost thawing and associated release of CO_2 (under aerobic conditions) and/or CH_4 (under anaerobic conditions)	~2.0°C	45(20-80) Gt C 0.09(0.04-0.16)°C; thawing already occurring at ~1.0°C	Estimated feedback by 2100	Estimates based on 48-50; observations from 51
Release of CH_4 from ocean methane hydrates[2]	~2.0-6.5°C (1000 - 5000 Gt C cumulative emissions)	Negligible by 2100	Gradual, slow release of C on millennial time scales to give +0.4-	52

			0.5°C feedback	
Weakening of land and ocean physiological C sinks that remove CO_2 from the atmosphere[3]	~2.0°C	Relative weakening of sinks by 0.25(0.13-0.37) °C	Estimated feedback by 2100	Rescaling of results from RCP4.5 "compatible emissions" scenario (53)
Increased bacterial respiration in the ocean, increasing release of CO_2 to the atmosphere	~2.0°C	12 Gt C 0.02°C	Estimated feedback by 2100	Rescaling of RCP8.5 results (54,55)
Amazon forest dieback, releasing CO_2 to the atmosphere, often through wildfire	2.0°C; tipping point possible in 3.0-5.0°C range	25(15-55) Gt C 0.05 (0.03-0.11) °C	Estimated feedback by 2100	Based on extrapolation of observed changes and model projections of dieback (56)
Boreal forest dieback, releasing CO_2 to the atmosphere, often through wildfire	2.0°C; tipping point possible in 3.0-5.0°C range	30(10-40) Gt C 0.06(0.02-0.10) °C	Estimated feedback by 2100	Based on extrapolation of observed changes and model projections (57-61)
Reduction of northern hemisphere spring snow cover, decreasing the albedo and thus amplifying regional warming	Scales with temperature increase	Contributes to polar amplification of temperature by factor of ~2	Fast – some reduction of snow cover already observed (63)	See Box 5.1 in 62
Arctic summer sea-ice loss, decreasing the albedo and thus amplifying regional warming	Current to RCP4.5; tipping point likely in 1.0-3.0°C range	Contributes to polar amplification of temperature by factor ~2	Fast – likely to have ice-free Arctic Ocean (summer) by 2040/50	Direct effect is largely regional but with global impacts on atmospheric and oceanic circulation; see Box 5.1 in 62
Antarctic summer sea ice loss, decreasing the albedo and amplifying regional warming	Recent observations of sea-ice loss at current levels of warming	Projected to be much smaller than feedback in northern hemisphere polar region (66)	Estimated loss of 30% by 2100	64,65 18,19
Polar ice sheet loss[4]	1.0-3.0°C	3-5 m sea-level rise from loss of West Antarctic Ice Sheet; up to 7 m from loss of Greenland Ice Sheet; up to 12 m from marine-grounded parts of East Antarctic Ice Sheet	Centuries to millennia	67-71

		.		

[1]Measured in most cases here as both an additional temperature rise (°C) by 2100 arising from the feedback and in Gt C emitted to the atmosphere. Where feedback strength was originally estimated in Gt C emitted to the atmosphere, the feedback strength was converted to °C assuming a 2°C temperature rise per 1000 GtC added to the atmosphere.
[2] The ocean methane hydrate feedback, although not likely to lead to a significant release of carbon in the 21st century, is included because on the longer term it is likely to be activated by a 2°C temperature rise, will lead to large releases of carbon, and is irreversible.
[3]Physiological carbon cycle feedbacks are already included in the calculation of a cumulative carbon budget consistent with a 2°C temperature rise (53)
[4]Focus here is on vulnerable parts of polar ice sheets, subject to the marine ice-sheet instability mechanism such as West Antarctica and parts of Greenland and East Antarctica, that could be lost in a timeframe up to 1000 years but whose tipping points could be crossed this century. Significant loss of polar ice sheets may affect not only sea-level rise but could also amplify long-term temperature rise by driving changes in ocean overturning circulation in the North Atlantic and Southern Ocean (67).

Permafrost thawing

The IPCC notes that there is *high confidence* that reductions in permafrost extent due to warming will cause thawing of some currently frozen carbon. However, there is *low confidence* in the magnitude of carbon losses through emissions of CO_2 and CH_4 to the atmosphere from this source. In the IPCC AR5 Ciais et al. (53) gave an estimate of 50 to 250 Gt C vulnerable to loss as both CO_2 and CH_4 between 2000 and 2100 under the high emissions RCP8.5 scenario.

Since the publication of the IPCC AR5, there have been several pertinent studies regarding permafrost thawing (48-50). These collectively give cumulative CO_2 and CH_4 emission figures for different RCP scenarios. For example, Schaefer et al. (48) give 27-100 Gt C release by 2100 under RCP4.5. Schneider von Deimling et al. (49) give 20-58 Gt C by 2100 under RCP2.6. Koven et al. (50) find a roughly linear loss of -14 to -19 Gt C/deg.C by 2100, that is, for 2°C warming 28-38 Gt C loss (which can then be multiplied by 1.1-1.18 to account for CH_4). Taking the latter estimate for 2°C warming in 2100, this corresponds to 28-38 Gt C loss as CO_2 or 31-45 Gt C as CO_2 equivalents including CH_4. We take 45 Gt C from this as our central estimate. We take 20 Gt C as our lower bound from the RCP2.6 study of (49), noting that RCP2.6 scenarios typically stabilize below 2°C warming. As an upper bound we scale down the upper estimate of Schaefer et al. (48) from RCP4.5 to ~80 Gt C (100*2/2.5).

Ocean bacterial respiration

The weakening of the ocean carbon pump involves different processes that diminish CO_2 drawdown in surface waters. The models used in Ciais et al (53) consider both abiotic (i.e., temperature effects on the solubility of CO_2 in seawater) and biotic processes. With respect to the latter, the models examine how expected changes in water column stratification brought about by surface ocean warming may impact ecosystem function. Thus, they assume that increased thermal stratification will lead to a reduction of nutrient input to the surface layer and a consequent decrease (*medium confidence*) in primary production which, in turn, will lead to a weakening of the biological pump whereby carbon is transported in sinking biological material from surface to deep ocean layers where it may be sequestered.

Another process that will impact carbon turnover in a warmer surface ocean and is not explicitly addressed in the models used by Ciais et al (53) is increased bacterial respiration in surface waters in response to increasing temperature (55). Increased remineralisation of organic material in surface waters through increased bacterial activity will lead to regeneration of nutrients (stimulating the production of some but not all phytoplankton groups) and an increase in pCO_2 (reducing the air to sea flux of CO_2).

Segschneider and Bendtsen (54) have examined the potential impact of temperature-dependent remineralisation in a warming ocean, including predicted changes in phytoplankton species composition (but not including potential effects of ocean acidification on calcifying organisms). On the basis of their analysis, we include here consideration of the changes in surface ocean bacterial activity. They estimated a ~18 Gt C cumulative sea-to-air flux under an RCP8.5 warming scenario, so this estimate must be adjusted for a 2°C global warming scenario. Sea surface temperatures increase less than the global average temperature; IPCC estimates that RCP4.5 leads to a 1°C warming of SST above present in 2060, and for RCP8.5 a 1.5°C warming of SST above present. Thus, for our analysis here, we estimate 2/3rds of the RCP8.5 effect i.e. ~12 Gt C.

Amazon dieback

We approach estimating potential carbon loss from the Amazon rainforest from (i) model projections of dieback (ii) extrapolation of observed changes.

The study of Jones et al. (56) suggests that at a global warming of 2°C the Amazon rainforest could be committed to a dieback of 40%. Dieback can take beyond the end of the century to fully unfold in their model. However, that model does not include fires, which could cause more much more rapid loss of carbon from the forest once it is under an unfavorable climate. To calculate the amount of carbon emissions associated with 40% dieback, we assume that the "dieback" is, in effect, a conversion of tropical forest to tropical savanna/grassland. We estimate a total C storage of 150-200 Gt for tropical forest in the Amazon basin (Table S3), which if replaced by tropical grassland/savanna would store 97-150 Gt C, a loss of 53-70 Gt C, 40% of which is 21-28 Gt C or ~25 Gt C.

Table S3: Carbon stored in main biomes

Biome	**C stored, Gt**	**Area of biome, $10^6 km^2$**	**Gt C/$10^6 km^2$**
Tropical/subtropical forest	547.8	20.33	26.94
Tropical/subtropical grassland/savanna	285.3	16.31	17.49
Desert/dry shrubland	178.0	25.22	7.06
Temperate grassland/savanna	183.7	14.50	12.67
Temperate forest	314.9	20.48	15.38
Boreal forest	348.2	19.23	18.11
Tundra	155.4	16.52	9.40

Sources:

Carbon stored per biome: GRID Arendal (http://old.grida.no/publications/rr/natural-fix/page/3725.aspx)

Area of biome: gk12glacier.bu.edu/wordpress/pasquarella/data-sets/location-and-area-of-biomes/

Observations indicate that the Amazon C sink decreased ~30% from 0.54 (0.45-0.63) Gt C/yr in the 1990s to 0.38 (0.28-0.49) Gt C/yr in the 2000s, partly due to increasing tree mortality linked to climate variability (72). Furthermore, individual drought events have reduced carbon uptake by 1.6 Gt C (2005) and 1.1 Gt C (2010) with the 2010 drought event making the Amazon essentially carbon neutral at 0.07 Gt C/yr (73). Extrapolation of the 0.16 Gt C/yr weakening of sink from the 1990s to 2000s for the remaining 83 years of this century gives a conservative minimum estimate of 13 Gt C equivalent emission from the Amazon, which we round to ~15 Gt C. Alternatively, assuming there continue to be on average two droughts per decade of comparable magnitude to the 2005 and 2010 events for the remaining eight decades of the century, the equivalent emission would be ~22 Gt C. This is in the range of the dieback estimates calculated above. Hence we consider ~25 Gt C to be a reasonable best estimate. Alternatively, if the observed rate of decline in Amazon C sink (0.016 Gt C/yr/yr) were to continue for the remainder of this century, then by the 2090s the Amazon would be a C source of 1.06 Gt C/yr and the equivalent carbon emission over the century would be ~80 Gt C, or for the 83 years from 2017 onwards the equivalent emission would be ~55 Gt C. We take the latter as an upper estimate.

Boreal forest dieback

We estimate potential carbon loss from the dieback of the boreal forest from (i) extrapolation of observed changes (ii) model projections of dieback.

1. Extrapolation from observed changes: Data from the Kurz and Apps (57) analysis of changes in the carbon budget for Canadian forests from 1920 to 1989 show an increase in carbon storage in biomass from 1920 to 1970 (11.0-16.4 Gt C), followed by a decrease from 1970 to 1989 (16.4-14.5 Gt C), driven largely by an increase in disturbance regimes (insect-induced stand mortality and fires) associated with a warming climate. DOM (Dead Organic Matter) showed an increase from 61.2 to 71.4 Gt C through the 70-year period. The increase in biomass and DOM over the 1920-1970 period is likely due, in part, to recovery from an earlier period of higher disturbance frequency.

If we assume that the rate of loss of biomass from 1970 to 1989 (1.9 Gt C through 20 years, or 0.095 Gt C per year) is maintained out to the end of the 21st century as the climate continues to warm, we estimate a loss of (0.095 Gt C/yr)(110 yrs) = 10.4 Gt C. Loss of about 10 Gt C over a century for the Canadian boreal forest is a reasonable (probably conservative) estimate when compared to an uptake in biomass of 5.4 Gt C through 50 years (from 1920 to 1970) as the forest recovered from an earlier high disturbance regime. Assuming that the Canadian boreal forest comprises about one-third of the global boreal forest, gives a crude estimate of ~30 Gt C, for the amount of carbon that could be lost by 2100 from boreal forest dieback. These estimates are for biomass only and do not include increased C release from DOM as disturbance regimes change. Nevertheless, we take this as our central estimate.

A different analysis finds a 73% reduction in the strength of the C sink in boreal regions from 0.152 Gt C/yr in previous decades to 0.041 Gt C/yr over 1997-2006 (58). Conservatively, if this 0.111 Gt C/yr reduction in sink strength persists over the subsequent 94 years to 2100 this corresponds to a C loss of 10.4 Gt C, in agreement with our Canada-only estimate above. Hence we take ~10 Gt C as our lower estimate. Alternatively, if the sink continues to decline at the observed rate 0.011 Gt C/yr/yr over 94 years this is equivalent to an emission of 49 Gt C, or for the 83 years from 2017 onwards the equivalent emission would be 38 Gt C. Hence we take ~40 Gt C as our upper estimate of the potential carbon loss through boreal forest dieback.

2. Model simulations of boreal forest conversion to savanna/grassland. The replacement of boreal forest at its southern boundary with steppe grasslands of lower carbon storage is consistent with grasslands being the neighbouring biome, as is an expansion of boreal forest into present-day tundra, which would increase carbon storage (59). Almost all models in the IPCC AR5 fail to capture the carbon dynamics of these biome transitions probably because they fail to appropriately capture disturbance factors (59). However, a novel estimate, based on 'climate vectors', of the equilibrium change in carbon storage expected under ~2°C warming (2040-2060 under RCP4.5) suggests a C loss over the 40-60°N region of 27 to 0.17 (mean12.3) Gt C, but with an increase in C storage of 3.7-16 (mean 7.2) Gt C for the boreal-forest transition (59).

Some earlier models also simulated boreal forest dieback (60,61), but both of these appear to simulate the ecosystem shifts at both the southern and northern edges of present-day boreal forests after they have reached equilibrium. Lucht et al. (60) simulated a replacement of the boreal forest at its southern border in a large stretch across Siberia and to the southwest of Hudson Bay in Canada under a relatively high warming scenario, but they do not quantify the area of boreal forest dieback or the carbon lost. Joos et al. (61) used a coupled climate-carbon cycle model to simulate changes in biome distribution and carbon dynamics using two emissions scenarios (moderate and high). For the moderate scenario, which gives a 2.8°C temperature rise by 2100, a large area of boreal forest in southern-central Siberia and a smaller area southwest of Hudson Bay in Canada are converted to savanna/woodland or grassland. This results in large losses of carbon in both areas (Plate 2 in their paper). Although Joos et al. (61) do not report the carbon storage changes by biome, a rough approximation can be made based on an estimated 50% loss of boreal forest to savanna/grassland. Then based on Table S2, the carbon lost to the atmosphere by conversion of half of the global boreal forest to temperate savanna/grassland is (19.23 10^6km^2)(0.5) x (18.11-12.67 Gt C/10^6km^2) = 52.3 Gt C. Assuming linear scaling with the level of temperature forcing would give an estimate of 52.3 Gt C (2.0/2.8) = 37.4 Gt C. This compares well with the upper estimate of 38 Gt C above based on Hayes et al. (58).

The Koven (59), Lucht et al. (60) and Joos et al. (61) studies all simulate an uptake of carbon in the far north as boreal forests migrate into present-day tundra regions. However, none of the modelling studies appear to capture the century-scale lag between the rapid loss of carbon from disturbance-driven dieback and the much slower migration and growth of boreal forests into the higher latitudes (e.g., 74). It is the pulse of carbon to the atmosphere resulting from this long lag time that we are estimating as the feedback resulting from boreal forest dieback (Table S1), an estimate we base on the observed carbon cycle dynamics of disturbance-driven dieback in the Canadian boreal forests. We emphasise that we are considering here the carbon dynamics of a highly transient phase of the Earth System (the potential 'Hothouse Earth' pathway) and one that is not close to equilibrium.

Critical biomes that support humanity (Section "What Is at Stake?" in the main text)

Table S4: Critical biosphere or Earth System features that support humanity

Biome or regional feature	**Importance for human well-being**	**Risks of a 'Hothouse Earth' pathway**
Productive agricultural regions – North America; West/central Europe; NE China; Indo-Gangetic Plain	Each of those regions provides food for >1 billion people or more	Depletion of soil fertility; changes in water availability; loss of coastal lands
Coral reefs	10% of marine fisheries, food source for 500 million people	Destruction of most reefs from warming and ocean acidification.
Tropical rainforests	Supports climate stabilization; centers of terrestrial biodiversity,	Amazon at risk from both climate and land-use change; SE Asia and Central Africa under increasing human pressure.
Tropical drylands	Support large human populations, particularly in Africa.	At risk of becoming too hot for humans; too hot and dry for agriculture.
Low-lying deltas and coastal regions	Centres of population, infrastructure, economic activity. Two-thirds of the world's mega-cities are less than 10 m above sea level.	Huge risks from coastal flooding to transport, infrastructure and coastal ecosystems. Economic damages could trigger regional or global economic collapse.
Regional monsoon systems	More than 1 billion people reliant on stability of South Asian Monsoon (SAM).	SAM vulnerable to high aerosol loading to warming Indian ocean and adjacent land.
Mountain glaciers	Provide freshwater for >1 billion people in Asia, South America	Melting at rapid rates, changing amount and timing of run-off
Riparian and wetlands	Provide freshwater for billions of people; carbon sequestration	loss of water in some places; increased flooding in others;

Human feedbacks in the Earth System (Section "Human Feedbacks in the Earth System" in the main text)

Table S5. Human actions that could steer the Earth System onto a 'Stabilized Earth' trajectory

Action	**Effect on Earth System (ES) trajectory**	**Time scale**	**Progress**
Enhancing or creating negative feedbacks through carbon sinks			
Forest and land management	Increase land carbon sinks by halting deforestation and increasing forests and other carbon sequestering land uses (75,76).	Decades	Rate of deforestation has slowed in many regions, but much more can be done
Soil management in agricultural zones	Better agricultural practices to replenish soil carbon lost in tillage could generate a significant below-ground carbon sink (77).	Decades to centuries	Better soil management not yet widespread so still remains a potential rather than actual C sink; questions about magnitude and potential reversibility of soil carbon sinks
Biodiversity conservation in land, coastal and marine systems	More biodiverse systems generally store more carbon than degraded ones, thus maintaining and enhancing biosphere sinks (77).	Centuries	Magnitude of effect is difficult to determine. In general, biodiversity loss and biosphere degradation continues in many regions.
Enhancing marine carbon uptake	Fertilization of ocean waters to stimulate uptake of CO_2 by marine phytoplankton	Decades to centuries	Uncertain effect as the magnitude of resulting carbon sequestration is unknown and eutrophication, possibly creating dead zones, is likely (78)
'Carbon Capture and Storage' (CCS)[1]	Remove CO_2 from the atmosphere and store it in geological formations. Often linked with bio-energy production (BECSS) (79).	Decades to centuries; storage time scale is millions of years, in principle	In R&D phase. Often energy intensive. Not currently economically feasible (but some facilities beginning to test the concept, e.g. a plant in Edmonton, Canada). BECSS, and other biomass energy production systems, likely to compete with agriculture and biodiversity conservation for land (80)
Reducing greenhouse gas emissions from fossil fuels and other sources			

[1]This carbon-sink feedback is fundamentally different from the others in that, in principle, it returns geological (fossil) carbon from the atmosphere to a geological formation, while the other sinks transfer carbon from one pool (atmosphere) to another (land or ocean) in the active carbon cycle. This carbon is vulnerable to return to the atmosphere whereas carbon stored safely in geological formations is not (81).

Replacement of fossil fuels with low or zero emission energy sources (82)	Reduces human impacts on radiative forcing and ocean acidification	Significant lags of decades to centuries. Further technological innovation necessary	Some progress has been made globally in reducing emissions in stationary energy (heat, electricity) (79, 83-86). Share of renewables in electricity sector doubling every 5-6 years. Costs of electricity storage are dropping; smart grids being developed. Electrification of transport is accelerating.
Reduce emissions from agriculture (livestock, rice, fertilizer)	In addition to CO_2, reduces methane and nitrous oxide emissions	Requires technological and behavioral change, e.g., low- or no-meat diets (87,88)	Emissions from livestock still increasing, some R&D
Reduction in production of chemicals such as HFCs (89,90)	Reduces concentration of chemicals with high GWP (Greenhouse Warming Potential)	Technological changes	Progress in eliminating HFCs
Reduce use of cement (91)	Reduces CO_2 emissions	Requires slowing of demand, CCS or use of alternative materials (e.g., wood, stone, carbon fiber)	No signs of slower growth although industry has made commitments to lower CO_2 intensity
Increased resource use efficiency in all human systems; reduce waste and use waste as a resource stream (recycling) (92,93)	Reduce the introduction of reactive N and P as well as novel substances into the biosphere and the rest of the Earth System; work towards.	Variable Technological innovation necessary, as well as change in consumer attitudes and business practices	Little progress globally but some examples emerging relating to recycling.
Modifying Earth's energy balance			
Reduce solar radiation input to Earth System through changing radiative or reflective properties of surface and atmosphere	Cooling effect as radiation is reflected or reduced	Decades	Concern about unintended consequences and technologies not yet economic or feasible at global scale; ocean acidification would continue (94-96)

Fundamental changes in societies			
Slow or reverse human population growth	Reduces demand on natural resources and overall consumption	Decades	Fertility rates are beginning to decline, mostly as a result of improved status of women, approaching replacement fertility at 2.4 children per woman. Lags and high fertility in some countries mean that although rates are slowing population growth will continue until at least mid-century.
Change consumption behavior	Reduce consumption of products associated with high GHG emissions and resource use — especially where consumption is far above average. For example, the wealthiest one billion people produce 60% of GHGs, whereas the poorest three billion produce only 5%.	Decades	Some consumer behavior change can be observed but rising incomes in many regions are increasing per capita consumption
Improved governance across scales, including the Earth System level	Develop policies to manage climate and Earth System in an equitable way. Policies include fiscal measures, incentives, regulation, information (including commitment to education of the general public) at all levels of government, business and organizations	Depends on strength of policy, number of actors, and implementation	Hundreds of international (UN) Conventions now exist to manage Earth System (e.g. climate, biodiversity, toxics). Paris agreement provides a measure of progress. Many non-nation state actors (e.g. cities, businesses) are committing to reduce emissions. The universal UN Sustainable Development Goals (SDGs) provide humanity with a first-ever roadmap integrating aspirational world development with Earth System sustainability. However, many conventions are disconnected from trade agreements; aviation and shipping emissions not covered in international agreements.

Value changes to support transformations to Earth System stewardship	Awareness of and education about Earth System vulnerability can contribute to changes in values that can change behavior, governance and intensity of innovation	Decades	Surveys indicate increased environmental awareness across many countries, decision makers and thought leaders (including popular culture and religion)
Technological innovation	Can provide new energy sources, food and transport, systems, efficiencies, carbon sinks, geoengineering, water reuse	Decades	Some low carbon and other technologies now spreading more rapidly (solar, genetics), others face opposition (solar, GM). Adoption often cost limited. R&D inadequate.
Management and governance philosophies emphasizing adaptiveness, complexity and uncertainty	Accepting and learning to live with uncertainty and change, and building resilience to Earth System change and for transformations	Decades	Influencing of key global actors has made some progress. Also, emergence of new global actors (e.g., development agencies, large businesses) that are actively working with a resilience focus

References: Supporting Information

1. Lenton TM (2015) *Earth System Science. A Very Short Introduction*. Oxford, Oxford University Press, 153 pp.
2. Clark PU, et al. (2016) Consequences of twenty-first-century policy for multi-millennial climate and sea-level change, *Nature Clim Change,* doi: 10.1038/nclimate2923.
3. Ruddiman WF, et al. (2016) Late Holocene Climate: Natural or Anthropogenic. Rev. Geophys. 54, doi:10.1002/2015RG000503.
4. Berger A, Loutre MF (2002) An exceptionally long interglacial ahead? *Science* 297: 1287–1288.
5. Steinhilber F, et al. (2012) 9,400 years of cosmic radiation and solar activity from ice cores and tree rings, *Proc. Nat. Acad. Sci.* 109:5967-5971.
6. Magny M (1993) Solar influences on Holocene climatic changes illustrated by correlations between past lake level fluctuations and the atmospheric ^{14}C record, *Quat. Res*. 40:1-9.
7. Magny M (2007) West-Central Europe, in *Lake Level Studies*, *Encyclopedia of Quaternary Science*, Elsevier,1389-1399.
8. Steffen W, Broadgate W, Deutsch L, Gaffney O, Ludwig C (2015), The trajectory of the Anthropocene: The Great Acceleration. *The Anthropocene Rev* 2:81-98.
9. Williams M, et al. (2015) The Anthropocene biosphere, *The Anthropocene Rev* 2:196-219.
10. McNeill JR, Engelke P (2016) *The Great Acceleration*. Harvard University Press.
11. IPCC (2013) Summary for Policymakers. 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, edited by Stocker TF, et al. Cambridge and New York, Cambridge University Press, pp 3-29.
12. NOAA (2016) State of the Climate: Global Analysis for Annual 2015. National Centers for Environmental Information, available at http://www.ncdc.noaa.gov/sotc/global/201513
13. Hawkins E, et al. (2017) Estimating changes in global temperature since the pre-industrial period. *Bull Am Meteor Soc* doi:10.1175/BAMS-D-16-0007.1
14. Lüthi D, et al. (2008) High-resolution carbon dioxide concentration record 650,000–800,000 years before present. *Nature* 453: 379-382.
15. Marcott SA, Shakun JD, Clark PU, Mix A (2013) A reconstruction of regional and global temperature for the past 11,300 years. *Science* 339:1198-1201.
16. Summerhayes CP (2015) *Earth's Climate Evolution*. Wiley, Chichester, 394pp.
17. Foster GL, Royer DL, Lunt DJ (2017) Future climate forcing potentially without precedent in the last 420 million years. *Nature Comms*, doi: 10.1038/nccomms14845
18. Sluijs A, et al. (2008) Eustatic variations during the Paleocene–Eocene greenhouse world, *Paleoceanography* 23: doi:10.1029/2008PA001615.
19. McInerney FA, Wing SL (2011) The Paleocene-Eocene Thermal Maximum – a perturbation of carbon cycle, climate, and biosphere with implications for the future. *Ann. Rev. Earth Planetary Sci.* 39:489-516.
20. Haywood AM, et al. (2011) Are there pre-Quaternary geological analogues for a future greenhouse gas-induced global warming? *Phil. Trans. Royal Soc. London*, *A*, 369:933-956.
21. Winguth AM, Thomas E, Winguth C (2012) Global decline in ocean ventilation, oxygenation, and productivity during the Paleocene-Eocene Thermal Maximum: Implications for the benthic extinction. *Geology* 40:263–266.
22. Zeebe RE, Ridgwell A, Zachos JC (2016) Anthropogenic carbon release rate unprecedented during the past 66 million years. *Nature Geosci* doi:10.1038/ngeo2681
23. Gutjahr M, et al., (2017) Very large release of mostly volcanic carbon during the Palaeocene-Eocene Thermal Maximum. *Nature* 548:573-577.
24. Le Quéré C, et al. (2017) Global Carbon Budget 2017. *Earth System Science Data Discussions* https://doi.org/10.5194/essdd-2017-123
25. Courtillot VE, Renne PR (2003) On the ages of flood basalt events. *Comptes Rendus Geoscience* 335:113-140.
26. Sahney S, Benton MJ (2008) Recovery from the most profound mass extinction of all time. *Proc. Biol. Sci*. 275**:**759-765.
27. IPCC (2013) Summary for Policymakers. 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, edited by Stocker TF, et al. Cambridge
28. Scheffers BR, et al. (2016) The broad footprint of climate change from genes to biomes to people. *Science* 354: doi: 10.1126/science.aaf7671
29. Diffenbaugh NS, Field CB (2013) Changes in ecologically critical terrestrial climate conditions. *Science* 341:486-492.
30. Rothman, DH (2017) Thresholds of catastrophe in the Earth system. *Science Advances* 3: doi 10.1126/sciadv1700906
31. Steffen W, et al. (2016) Stratigraphic and Earth System approaches to defining the Anthropocene. *Earth's Future* 4: doi: eft2/2016EF000379
32. Williams M, et al. (2015) The Anthropocene biosphere, *The Anthropocene Rev* 2:196-219.
33. Hooke RLeB, Martin-Duque JF, Pedraza J (2012) Land transformation by humans. A review. *GSA Today* 22: doi: 10.1130/GSAT151A.a
34. Ellis EC, et al. (2010) Anthropogenic transformation of the biomes, 1700 to 2000. *Global Ecol. Biogeogr.* 19:589–606.

35. Ceballos G, et al. (2015) Accelerated modern human–induced species losses: Entering the sixth mass extinction, *Science Adv* 1:e1400253, DOI: 10.1126/sciadv.1400253.
36. McNeely J (2001) Invasive species: A costly catastrophe for native biodiversity, *Land Use and Water Resour. Res.* 1:1-10.
37. NOAA (2017) Global Analysis - Annual 2016. Accessed at: https://www.ncdc.noaa.gov/sotc/global/201613.
38. Kopp, RE, Simons FJ, Mitrovica JX, Maloof AC, Oppenheimer M (2009) Probabilistic assessment of sea level during the last interglacial stage. *Nature* 462:863-868.
39. Miller KG, et al. (2012), High tide of the warm Pliocene: implications of global sea level for Antarctic deglaciation. *Geology* 40:407-410.
40. Greenop R, Foster GL, Wilson PA, Lear CH (2014) Middle Miocene climate instability associated with high-amplitude CO_2 variability. *Paleoceanogr.* 29: doi: 10.102/2014PA002653
41. Kominz MA, et al. (2008) Late Cretaceous to Miocene sea-level estimates from the New Jersey and Delaware coastal plain coreholes: an error analysis. *Basin Research* 20:211-226.
42. Ganopolski A, Winkelmann R, Schellnhuber HJ (2016) Critical insolation–CO_2 relation for diagnosing past and future glacial inception. *Nature.* 529**:**200-203.
43. Berger A, Loutre MF (2002) An exceptionally long interglacial ahead? *Science* 297: 1287–1288.
44. Waters CN, et al. (2016) The Anthropocene is functionally and stratigraphically distinct from the Holocene, *Science* 351**:**6269.
45. Archer D, Buffett B, Brovkin V (2009) Ocean methane hydrates as a slow tipping point in the global carbon cycle. *Proc Natl Acad Sci (USA)* 106: 20596-20601.
46. Solomon S, Plattner G-K, Knutti R, Friedlingstein P (2009) Irreversible climate change due to carbon dioxide emissions. *Proc Natl Acad Sci USA* 106:1704-1709.
47. Yin Q, Berger A (2012) Individual contribution of insolation and CO_2 to the interglacial climates of the past 800,000 years. *Clim Dynamics* 38:709-724.
48. Schaefer K, Lantuit H, Romanovsky VE, Schuur EAG, Witt R (2014) The impact of the permafrost carbon feedback on global climate. *Environ Res Lett* 9:085003.
49. Schneider von Deimling T, et al. (2105) Observation-based modelling of permafrost carbon fluxes with accounting for deep carbon deposits and thermokarst activity. *Biogeosciences* 12: 3469-3488.
50. Koven CD, et al. (2015) A simplified, data-constrained approach to estimate the permafrost carbon-climate feedback. *Phil Trans R Soc A* 373:20140423, http://dx.doi.org/10.1098/rsta.2014.0423
51. Chadburn SE, Cox PM, Friedlingstein P, Hugelius G, Westermann S (2017) An observation-based constraint on permafrost loss as a function of global warming. *Nature Climate Change* 7:340-344 doi:10.1038/NCLIMATE3262
52. Archer D, Buffett B, Brovkin V (2009) Ocean methane hydrates as a slow tipping point in the global carbon cycle. *Proc Natl Acad Sci (USA)* 106: 20596-20601.
53. Ciais P, et al. (2013), Carbon and other biogeochemical cycles, 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, edited by Stocker TF, et al., Cambridge and New York, Cambridge University Press, pp465–570.
54. Segschneider J, Bendtsen J (2013) Temperature-dependent remineralization in a warming ocean increases surface pCO_2 through changes in marine ecosystem composition. *Global Biogeochem Cycles* 27:1-12.
55. Bendtsen J, Hilligsøe KM, Hansen J, Richardson K (2015) Analysis of remineralisation, lability, temperature sensitivity and structural composition of organic matter from the upper ocean. *Prog Oceanogr* 130:125-145.

56. Jones C, Lowe J, Liddicoat S, Betts R (2009) Committed terrestrial ecosystem changes due to climate change. *Nature Geosci.* 2:484-487.
57. Kurz WA, Apps MJ (1999) A 70-year retrospective analysis of carbon fluxes in the Canadian forest sector. *Ecol Applications* 9:526-547.
58. Hayes D J, et al. (2011) Is the northern high-latitude land-based CO_2 sink weakening? *Global Biogeochem Cycles* 25: GB3018, doi:10.1029/2010GB003813
59. Koven CD (2013) Boreal carbon loss due to poleward shift in low-carbon ecosystems. *Nature Geoscience* 6:452-456.
60. 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: 6.
61. Joos F., et al. (2001) Global warming feedbacks on terrestrial carbon uptake under the Intergovernmental Panel on Climate Change (IPCC) emission scenarios. *Global Biogeochem Cycles* 15:891-907.
62. Masson-Delmotte V, et al. (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.* edited by Stocker TF, et al. Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA, pp383-464.
63. Vaughan DG, et al. (2013) Observations: Cryosphere. 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* [Stocker TF, et al. (eds.)]. Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA.
64. Turner J, et al. (2009) Non-annular atmospheric circulation change induced by stratospheric ozone depletion and its role in the recent increase of Antarctic sea ice extent. *Geosphys Res Lett* doi: 10.1029/2009GL037524
65. NASA (2017) https://www.nasa.gov/feature/goddard/2017/sea-ice-extent-sinks-to-record-lows-at-both-poles
66. Kopp, RE, Simons FJ, Mitrovica JX, Maloof AC, Oppenheimer M (2009) Probabilistic assessment of sea level during the last interglacial stage. *Nature* 462:863-868.
67. Hansen J, et al. (2016) Ice melt, sea level rise and superstorms: evidence from paleoclimatedata, climate modeling, and modern observations that 2°C global warming could be dangerous. *Atmos. Chem. Phys*. 16: 3761–3812.
68. Pollard D, DeConto RM (2009) Modelling West Antarctic ice sheet growth and collapse through the past five million years. *Nature* 458:329-332.
69. Pollard D, DeConto RM, Alley RB (2015) Potential Antarctic Ice Sheet retreat driven by hydrofracturing and ice cliff failure. *Earth Plan. Sci. Lett*. 412:112-121.
70. DeConto RM, Pollard D (2016) Contribution of Antarctica to past and future sea-level rise. *Nature* 531:591-597.
71. Rintoul SR, et al. (2016) Ocean heat drives rapid basal melt of the Totten Ice Shelf. *Science Advances* 2:e1601610.
72. Brienen RJW, et al. (2015) Long-term decline of the Amazon carbon sink. *Nature* 519: 344-348.
73. Feldpausch TR, et al. (2016) Amazon forest response to repeated droughts. *Global Biogeochem Cycles* 30: 964-982.
74. Smith TM and Shugart HH (1993) The transient response of terrestrial carbon storage to a perturbed climate. Nature 361: 523-525.
75. Boucher D, Roquemore S, Fitzhugh E (2013) Brazil's success in reducing deforestation. *Tropical Conservation Science* 6:426-445.

76. Smith P, et al. (2014) Agriculture, Forestry and Other Land Use (AFOLU). In: *Climate Change 2014: Mitigation of Climate Change. Contribution of Working Group III to the Fifth Assessment Report of the IPCC*, edited by Edenhofer O, et al. Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA, pp811-922.
77. Strassburg BBN, et al. (2009) Global congruence of carbon storage and biodiversity in terrestrial ecosystems. *Conservation Letters* **3**:98-105.
78. Royal Society (2009). Geoengineering the climate: science, governance and uncertainty. The Royal Society, UK, 82pp.
79. Rockström J, et al. (2017) A roadmap for rapid decarbonization. *Science* 355:1269-1271.
80. Boysen LR, et al. (2017) The limits to global-warming mitigation by terrestrial carbon removal. *Earth's Future* doi: 10.1002/2016EF000469
81. Mackey B, et al. (2013) Untangling the confusion around land carbon science and climate change mitigation policy. *Nature Climate Change* 3: 552-557.
82. Climate Change 2014: *Mitigation of Climate Change. Contribution of Working Group III to the Fifth Assessment Report of the Intergovernmental Panel on Climate Change* [Edenhofer O, et al. (eds.)]. Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA.
83. Schellnhuber HJ, Rahmstorf S, Winkelmann R (2016) Why the right climate target was agreed in Paris. *Nature Climate Change*, 6:649-653.
84. Jackson RB, et al. (2015) Reaching peak emissions. *Nature Climate Change*, http://dx.doi.org/10.1038/nclimate2892.
85. IEA (International Energy Agency) (2016) Decoupling of Global Emissions and Economic Growth Confirmed.
Retrieved June 15, 2016 (http://www.iea.org/newsroomandevents/pressreleases/2016/march/decoupling-of-global-emissions-and-economic-growth-confirmed.html). Google Scholar
86. Fischer-Kowalski M (ed) (2011) *Decoupling Natural Resource Use and Environmental Impacts from Economic Growth*. New York NY USA: UNEP Publishing.
87. Bennetzen EH, Smith P, Porter JR (2016) Decoupling of greenhouse gas emissions from global agricultural production: 1970-2050. *Global Change Biol*. 22:763-781.
88. Herrero M, et al. (2016) Greenhouse gas mitigation potentials in the livestock sector. *Nature Climate Change* 6:452-461.
89. Xu, Yangyang, Durwood Zaelke, Guss JM Velders, and Veerabhadran Ramanathan. "The role of HFCs in mitigating 21st century climate change." *Atmospheric Chemistry and Physics* 13, no. 12 (2013): 6083-6089.
90. Fischedick M, et al. (2014) Industry. In: *Climate Change 2014: Mitigation of Climate Change. Contribution of Working Group III to the Fifth Assessment Report of the Intergovernmental Panel on Climate Change* [Edenhofer O, et al. (eds.)]. Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA.
91. Rogelj, J., Meinshausen, M., Schaeffer, M., Knutti, R. and Riahi, K., 2015. Impact of short-lived non-CO_2 mitigation on carbon budgets for stabilizing global warming. *Environ Res Lett*, *10*(7), p.075001.
92. Daly HE, Cobb JB (1989) *For the common good: Redirecting the economy toward community, the environment, and a sustainable future.* Beacon Press, Boston MA USA.
93. Jackson T (2009) *Prosperity without Growth: economics for a finite planet*. London: Routledge (Earthscan).
94. Crutzen PJ (2006) Albedo enhancement by stratospheric sulfur injections: A contribution to resolve a policy dilemma? *Clim Change* 77: 211. doi:10.1007/s10584-006-9101-y
95. Barrett S, et al. (2014) Climate engineering reconsidered. *Nature Clim Change* 4:527-529.

96. Mathesius S, Hofmann M, Calderia K, Schellnhuber HJ (2015) Long-term response of oceans to CO_2 removal from the atmosphere. *Nature Climate Change* 5:1107-1113.