Climate updates

What have we learnt since the IPCC 5th Assessment Report?

REFERENCES

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Contents

Introduction

This document provides information supplementary to the Climate Updates report. Each section is the partner to a ‘question’ in the main report and provides background information to the conclusions reached there as well as a limited bibliography.

How sensitive is global temperature to increasing greenhouse gases?

SUPPLEMENTARY INFORMATION

The IPCC quotes used in this section comes from IPCC AR5, WGI, Summary for Policymakers, section D.2, page 14.

Note that although climate sensitivity values are defined in terms of carbon dioxide doubling, these typically (and specifically in the IPCC reviews) encompass the impact of all human factors, including other greenhouse gases such as methane, with their radiative impacts scaled to a 'carbon dioxide equivalent'.

A selection of papers contributing to new results or conclusions follows:
Reviews of the methods used in calculating equilibrium climate sensitivity and transient climate response and of recent developments in understanding the disparities: *Forster* (2016), *Knutti et al.* (2017).

Results based on energy balance models: *Bates* (2016), *Lewis* (2016), *Skeie et al.* (2014).

Evidence that climate sensitivity depends on background state: *Friedrich et al.* (2016), *He et al.* (2017), *Kohler et al.* (2015).

Regional variations in the temperature response affect derived sensitivity values: *Armour et al.* (2013), *Frolicher et al.* (2014), *Gregory and Andrews* (2016)

Climate sensitivities derived from GCMs underestimate the long term effect: *Armour* (2017), *Proistosescu and Huybers* (2017).

Impact of spatial heterogeneity in forcings on sensitivity estimates: *Kummer and Dessler* (2014), *Marvel et al.* (2016), *Shindell et al.* (2015).

Bias introduced by geographical distribution of measurements: *Richardson et al.* (2016).

Assessment of GCMs based on realism of cloud simulation: *Brient and Schneider* (2016), *Stevens et al.* (2016), *Tan et al.* (2016), *Zhai et al.* (2015).

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Bates J. 2016 Estimating climate sensitivity using two-zone energy balance models. *Earth and Space Science* **3**, 207-225. (doi:10.1002/2015ea000154)

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Frolicher T, Winton M, and Sarmiento J. 2014 Continued global warming after CO2 emissions stoppage. *Nat Clim Change* **4**, 40-44. (doi:10.1038/nclimate2060)

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He J, Winton M, Vecchi G, Jia L, and Rugenstein M. 2017 Transient climate sensitivity depends on base climate ocean circulation. *J Climate* **30**, 1493-1504. (doi:10.1175/jcli-d-16-0581.1)

Knutti R, Rugenstein MAA, and Hegerl GC. 2017 Beyond equilibrium climate sensitivity. *Nature Geoscience* 10, 727–736. (doi:10.1038/ngeo3017).

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Kummer J, Dessler A. 2014 The impact of forcing efficacy on the equilibrium climate sensitivity. *Geophys Res Lett* **41**, 3565-3568. (doi:10.1002/2014gl060046)

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Zhai C, Jiang J, and Su H. 2015 Long-term cloud change imprinted in seasonal cloud variation: More evidence of high climate sensitivity. *Geophys Res Lett* **42**, 8729-8737. (doi:10.1002/2015gl065911)

How are methane concentrations changing and what does this mean for the climate?

SUPPLEMENTARY INFORMATION

The IPCC quotes used in this section come from IPCC AR5, WGI, Chapter 2, section 2.2.1.1.4 and IPCC AR5, WGI, TS.2.8.3 respectively.

The value of “about 55%” for the warming effect of methane emissions relative to CO_2 emissions, for the period 1750 – 2011, originates from IPCC AR5 WG1, Figure SPM.5, page 14 (see also Figure TS.7, page 57). This value includes the effect of methane emissions on methane itself, and on concentrations of ozone, stratospheric water vapour and (via oxidation of methane) CO_2.

Global methane trends are regularly updated by the US National Oceanic and Atmospheric Administration’s Global Monitoring Division at its Earth System Research Laboratory: https://www.esrl.noaa.gov/gmd/ccgg/trends_ch4/

Further details on these observations and their analysis can be found in Dlugokencky *et al*. (1994) and Masarie and Tans (1995).

Several recent papers discuss the recent trends in methane concentrations, and its isotopic composition, and the possible drivers of these trends, particularly focusing on changes in emissions. These include Nisbet *et al*. (2016), Saunois *et al*. (2016a, 2016b), Schaefer *et al*. (2016) and Schwietzke *et al*. (2016)

Rigby *et al*. (2017) and Turner *et al*. (2017) particularly discuss the possible role of changes in atmospheric oxidation in explaining recent methane trends, while Prather and Holmes (2017) provide a commentary on the implications of these two papers and their relation to other recent work on the causes of methane trends, some of which are mentioned above.

There has been a recent indication of the possible need for a revision of calculations of the effect of changes in methane on the Earth’s radiation budget (Etminan *et al*. 2016). The direct effect of methane (i.e. excluding methane’s impact on ozone etc) would be increased by 25% relative to the values given in AR5. However, at the time of writing this result had not been corroborated by other studies and so it is not included in the main text.

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Turner A, Frankenberg C, Wennberg P, Jacob D. 2017 Ambiguity in the causes for decadal trends in atmospheric methane and hydroxyl. *P NATL ACAD SCI USA* **114**, 5367-5372. (doi:10.1073/P Natl Acad Sci USA.1616020114)

Was there a "pause" in global warming?

SUPPLEMENTARY INFORMATION

The IPCC quotes used in this section comes from IPCC AR5, WGI, Summary for Policymakers, Section B.1.

The AR5 also concluded that the reduced surface warming trend during 1998 – 2012 was "due in roughly equal measure to a reduced trend in radiative forcing and a cooling contribution from natural internal variability" and (as quoted in the main text) that "The reduced trend in radiative forcing is primarily due to volcanic eruptions and the timing of the downward phase of the 11-year solar cycle." With regard to the comparison between models and observations AR5 said: "There is medium confidence that natural internal decadal variability causes to a substantial degree the difference between observations and the simulations"; "There may also be a contribution from forcing inadequacies and, in some models, an overestimate of the response to increasing greenhouse gas and other anthropogenic forcing (dominated by the effects of aerosols)." See IPCC AR5, WGI, Summary for Policymakers, Section D.1 for full statements.

Global surface temperature observations

Observation-based estimates of global surface air temperature continue to be refined and improved. In particular, understanding of biases between different measurement types, particularly of sea surface temperatures, has improved during the period of discussion about the hiatus (e.g. Morice *et al.*, 2012) and since AR5 (e.g. Huang *et al.* 2015). Cowtan & Way (2014) showed that the treatment of regions where no observations are available (i.e. data gaps) is an important issue. They showed that two methods to infill gaps, particularly over the Arctic, resulted in a larger trend in surface warming over the hiatus period than was suggested by the data with no infilling. Overall, the relative slowdown in global temperature rise during the hiatus period is now less pronounced than in earlier datasets (Medhaug *et al.*, 2017). These datasets will continue to be refined in future. Global surface air temperatures in both 2015 and 2016 were likely more than 1°C above pre-industrial levels (Hawkins *et al.* 2017).

Earth's energy budget

Efforts to better understand the energy imbalance at the top of the atmosphere have demonstrated that the climate system continued to accumulate heat during the 'hiatus' period (e.g. Allan *et al.*, 2014, Xie & Kosaka 2017). Ocean heat content has also continued to increase (Abraham *et al.*, 2013) and global sea levels have also continued to rise (see sea level topic).

Comparison between models and observations

Several factors are now understood to affect any comparison of global surface temperature observations with model simulations. Evidence suggests that a series of small volcanic eruptions and a downturn in solar activity were a small cooling influence on the real world, but neither were represented in the CMIP5 simulations used in AR5 (Schmidt *et al.*, 2014, Medhaug *et al.*, 2017). In addition, the observations are an incomplete blend of sea surface temperatures over the ocean and air temperatures over the land, whereas the simulations produce geographically complete surface air temperatures. When comparing trends, AR5 accounted for some but not all of these differences. Their effect has now been better quantified (Cowtan *et al*., 2015; Richardson *et al*., 2016) and Medhaug *et al*. (2017) conclude that the observations and simulations can be reconciled by considering these two factors and also internal variability associated with cooler sea surface temperatures in the Pacific (e.g. Kosaka & Xie, 2013). These three factors are roughly equally important for resolving the apparent discrepancy. Different choices about the appropriate 'baseline' from which to measure changes in temperature can also affect the size of any apparent discrepancy between simulations and observations, but this has received less attention (Hawkins & Sutton, 2016).

Satellite-based temperature estimates

Estimates of global atmospheric temperature are available from interpreting satellite measurements of atmospheric microwave emissions since 1979. These datasets have larger uncertainties than the surface temperature observations due to complex issues around combining data from the small number of satellites used, and are being continually refined and developed. Improved understanding of biases in these satellite-based temperature estimates has shown that the rate of atmospheric warming during the period 1998 – 2012 was greater than the evidence available at the time of AR5 suggested (Wentz & Mears, 2016; Mears & Wentz, 2017). 2016 was also the warmest year in these revised datasets. An apparent inconsistency between observations and simulations of atmospheric warming rates was discussed by Santer *et al*. (2017), who concluded that these differences were likely due to the combined effects of two factors: deficiencies in the natural radiative forcings in the simulations and natural internal variability. There was little evidence that the differences were related to a large error in the sensitivity of climate models to greenhouse gas increases.

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How high could sea level rise because of anthropogenic climate change?

SUPPLEMENTARY INFORMATION

The IPCC quotes used in this section comes from IPCC WG1 AR5, Summary for Policymakers, Section E6.

Observed and projected sea level rise are not globally uniform, as discussed in the IPCC WG1 AR5 chapter 13, on sea level change. As the climate becomes warmer, more heat is stored in some parts of the ocean than others, and sea level rise due to thermal expansion of sea water is generally greater in those regions. Changes in wind-driven currents, and in salinity (due to changes in precipitation, evaporation and river inflow), also affect local sea level. In addition, loss of mass from an ice sheet means its gravitational attraction of the nearby ocean is reduced, the rotation of the Earth is affected, and the crust rises under the ice-sheet as weight is removed, with the surprising consequence that sea level falls in regions close to ice sheets which are losing mass, while sea level rise is enhanced further away. For all these reasons, one must remember that global mean and local sea level rise are not equal, although about 70% of the global coastlines are projected to experience a sea level change within 20% of the global mean sea level rise. Estimates of the number of people at risk from coastal flooding show that a rise of even 50 cm exposes tens of millions of additional people to flooding each year (Brown *et al*., 2016).

Attribution of the anthropogenic influence on sea level change shows that it has been the dominant influence on sea level rise since 1970 (Slangen *et al*., 2016). Natural factors were the cause of the majority of the global glacier mass loss since the mid-19th century, but in recent decades anthropogenic influence has become more important (Marzeion *et al*., 2014). Improved calculations of the components contributing to sea level change are consistent (Chen *et al*., 2017; Dieng *et al*., 2017) with the observations from satellite altimetry after they have been corrected (Watson *et al*., 2015) for a small systematic drift during the early years of deployment. The data show an acceleration in sea level rise in the latter part of the satellite era (Chen *et al*., 2017; Dieng *et al*., 2017), which is consistent with the largest added component coming from the Greenland Ice Sheet.

The possible onset of marine ice sheet instability at Thwaites Glacier (West Antarctica) was most conspicuously raised in Joughin *et al*. (2014). A prominent modelling study that includes some new mechanisms proposed to induce instability of ice shelves (DeConto and Pollard, 2016) has projected much larger and faster retreat of Antarctic ice than previous work. In particular under RCP8.5 it gives more than a metre of sea level rise from Antarctica alone by 2100, and 15 m by 2500 (implying significant loss of East Antarctic ice). However this is an outlier so far: much lower contributions by 2100 from parts or the whole of the Antarctic ice sheet have been produced in other models (Cornford *et al*., 2015; Favier *et al*., 2014; Golledge *et al*., 2015; Joughin *et al*., 2014; Levermann *et al*., 2014; Ritz *et al*., 2015), consistent with AR5 assumptions about the possible development of marine ice sheet instability. These studies therefore increase confidence in the AR5 likely range (Clark *et al*., 2015), while not excluding higher possibilities. Beyond 2100 and on the longer term, sea level rise due to loss of Antarctic ice is very uncertain and could be much greater. For example, one paper that simulates the long-term prognosis for Antarctic ice (Golledge *et al*., 2015) gives as little as 0.1 m sea level rise from Antarctica by 2100 under RCP8.5, although the temperature rise causes a commitment to loss of most of West Antarctic Ice Sheet (with a 5 m sea level rise) by the year 5000. Another study (Clark *et al*., 2016) argues that the consequences of policy decisions in this century for the next 10,000 years should be considered; they estimate that CO_2 already emitted will lead to an eventual sea level rise of about 1.7 m, but if a similar amount is emitted again, there will be a further committed rise of about 9 m, mostly from Antarctica.

There is increasing evidence, summarised in a review (Dutton *et al*., 2015), that sea level was 6 – 9 m higher in the last interglacial. Reconstructions of polar temperatures imply that this was under polar atmospheric and ocean temperatures within the range of those projected for 2100 (Capron *et al*., 2014).

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Watson C, White N, Church J, King M, Burgette R, Legresy B. 2015 Unabated global mean sea-level rise over the satellite altimeter era. *Nat Clim Change* **5**, 565-568. (doi:10.1038/nclimate2635)

Decreasing Arctic sea ice – is there any influence on the weather in middle latitudes?

SUPPLEMENTARY INFORMATION

The IPCC quotes used in this section come from IPCC AR5, WG1, Summary for Policymakers, section B.3.

Mori *et al*. (2014) showed, using both reanalysis data and atmospheric model simulations, that sea-ice decline leads to more frequent Eurasian blocking situations, which in turn favour cold-air advection to Eurasia, and hence severe winters in that region.

Screen *et al*. (2015) show that very cold winters have become less frequent in the 21st century. 2013/14 was not extreme in the longer record, though it was for recent winters. Their climate model results find that Arctic ice loss lessens the chance of North American cold extremes.

Horton *et al*. (2015) analyse observations and find an increase in wintertime low temperature extremes over central Asia since 1990 — the period over which the extent of Arctic sea ice has rapidly declined. They attribute this increase primarily to the increased occurrence of a circulation pattern associated with the transport of cold Arctic air into central Asia.

Overland *et al*. (2015) give a review of the area and stress its complexity. They suggest that reductions in sea ice in the Barents-Kara (B-K) Seas can lead to a wave pattern in the pressure field that can influence Ural blocking, and that North American cold can result from impacts on Greenland blocking.

Petrie *et al*. (2015) perform experiments with coupled ocean models that indicate that Arctic sea loss leads to much higher temperatures there and that this warming spreads into middle latitudes. They find a weak response in the North Atlantic Oscillation pattern in summer, but not in winter.

A perspective by Overland *et al*. (2016) stresses the difficulties in determining the possible impact on middle latitude weather of Arctic sea ice loss due to short time series, small signal to noise and other possible influences on the weather. It recognises that some case studies may give evidence of linkages.

Kretschmer *et al*. (2016) applied the method of causal effect networks to reanalyses and found that a loss of Barents/Kara sea ice can indeed be considered a causal driver of a weakened tropospheric polar vortex.

Ruggieri *et al*. (2017) propose a sequence of processes by which sea ice loss in the B-K sea may later impact on a particular pattern of variability in the North Atlantic region. This involves a local response followed by interaction with weather systems and later impacts via the stratosphere.

The analysis of Luo *et al*. (2017) suggests that weak westerly flow associated with Arctic amplification can encourage the occurrence of quasi-stationary Ural blocking that leads to intense winter cold in Eurasia.

Vavrus *et al*. (2017) test the hypothesis that Arctic Amplification leads to a weaker westerly wind with more meanders using 40 simulations of the past and future using a climate model. They find evidence that the weaker wavier flow can enhance summer heat over North America and perhaps cold troughs there in winter.

An observational analysis by Chen *et al*., (2017) suggests that the westward movement of Greenland blocks is related to weaker westerly winds and that these lead to colder winter weather over North America.

According to Screen (2017), regional sea ice loss generally gives a local thermodynamic response, but in some regions, e.g. the B-K Seas, there can be a dynamic response giving circulation changes leading to some areas being warmer and others cooler. The response to sea ice loss around the Arctic is not equal to the sum of the responses from the different regions.

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Have temperature and rainfall extremes changed and how will they change in the future?

SUPPLEMENTARY INFORMATION

The IPCC quotes used in this section comes from IPCC AR5, Summary for Policymakers, D3, B1, E1 and E2.

Trends in the frequency and intensity of many extremes have continued (e.g. Cassou and Cattiaux, 2016; Fischer and Knutti, 2015); and, for example, heat waves continued to increase in frequency during the 'hiatus' (Seneviratne *et al*., 2014). The importance of soil moisture feedback has been more strongly recognized, with dry conditions intensifying heat waves (e.g. Miralles *et al*., 2014) and spring moisture causing summer heat waves in some regions to occur earlier, and be hotter, (e.g. Cowan *et al*., 2017).

Since AR5, analysis of specific extreme events has continued to robustly indicate that humans have made some individual heat waves more likely, and cold spells less likely (NAS, 2016; Stott *et al*., 2016); methods to quantify this change are based on analysis of change in observations, and on climate modelling. For temperature-related extremes, results tend to indicate robust human influences in most cases (see Herring *et al*., 2016). For example, in the case of the 2015 summer heat wave in Europe, both Sippel *et al*. (2016) and Dong *et al*. (2016) find a human contribution to the heat wave. The estimated risk ratio indicates that the probability of such an event increased due to human influences, although quantitatively, the risk ratio varies with method and location.

For temperature and temperature-related variables, the changes directly related to global warming have a strong influence on regional climate change including extreme events. A moister warmer atmosphere is more conducive to heavy rainfall (e.g Pfahl *et al*., 2017). Changes in atmospheric circulation are presently poorly understood and difficult to predict and project, and such changes may moderate some changes in extremes that are expected directly from warming. This is particularly important for rainfall extremes, droughts or convective events that are also more strongly influenced by natural climate variability. This uncertainty remains a limiting factor in confidence in event attribution, particularly for event types that are not as strongly influenced by warming (see NAS, 2016; Trenberth *et al*., 2015; Shepherd, 2016; Vautard *et al*., 2016). Results can also vary depending if a study considers how extreme the rainfall would have been without greenhouse gas increases for the same synoptic situation, or if it considers how extreme rainfall frequency overall has changed in a region (NAS, 2016).

Analysis of projections of future climate change continues to indicate a tendency towards more and hotter hot extremes and reduced cold extremes, and increase in heavy precipitation in many regions (e.g. Fischer *et al*., 2013a,b), and there is some indication that future precipitation extreme changes could be larger than anticipated (e.g. Kendon *et al*. 2014, Borodina *et al*., 2017) for example, because of the inability of climate models to adequately resolve convective precipitation. This is consistent with observed past changes showing slightly larger changes than observed, as discussed in AR5 (see also Zhang *et al*., 2013).

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Are there thresholds beyond which particularly dangerous or irreversible changes may occur?

SUPPLEMENTARY INFORMATION

The IPCC quotes used in this section comes from the Thematic Focus Element TFE.5 of WG1, box entitled "Irreversibility and abrupt change".

Rapid sea level change due to collapse of the Greenland or West Antarctic Ice Sheets is not considered under this question because it is the focus of another question. Other thresholds have been postulated, and are also not considered here.

Atlantic Meridional Overturning Circulation (AMOC): New measurements of the palaeoceanographic proxies Pa/Th and $\delta^{13}C$ (Henry *et al*., 2016) suggest that changes in ocean circulation did accompany the rapid climate changes of the period from 60 – 25 ka ago, although proxies (Pa/Th and ε_{Nd}) do not suggest complete collapse during each cold stage (Bohm *et al*., 2015). Along with evidence about the phasing of Antarctic and Greenland temperature change (WAIS Divide Project Members, 2015), this means that the causal chain linking past rapid climate change and AMOC change is better-supported. Studies using expanded observations of the AMOC from the RAPID-MOCHA array (McCarthy *et al*., 2015) and from other observation methods (Buckley and Marshall, 2016) show how variable the AMOC is on different timescales, and discuss the potential for decadal forecasting. There has been a reduction in AMOC strength since the RAPID-MOCHA array was deployed, but the variability means that it is not yet possible to say whether this is significant. Model bias can inhibit the possibility of a bistable AMOC (one that has a strong and weak or off state) (Mecking *et al*., 2017). Recent work (Liu *et al*., 2017) suggests that when model biases are corrected collapse may occur, implying that models may be too stable against AMOC change. Although current models do not produce a sudden threshold AMOC collapse, a model intercomparison (Bakker *et al*., 2016) does find substantial but gradual weakening of the AMOC, by a median of 37% by 2100 under RCP8.5 and 74% by 2300.

Permafrost: A review published in 2015 (Schuur *et al*., 2015) made new estimates of the amount of carbon stored in permafrost, both near the surface and at depth. It also summarised experimental evidence about how much carbon would be released, and in what form, when permafrost melts. This led to a new estimate that about 100 Pg of cumulative carbon emissions (with a wide uncertainty) would be released from thawing permafrost by 2100 under RCP8.5. This leads to a significant positive feedback, but the review emphasised that emissions are "likely to be gradual and sustained rather than abrupt and massive". A recent modelling study estimated that permafrost carbon releases could contribute up to 12% of the change in global mean temperature by 2100 (Burke *et al*., 2017). Studies since 2013 therefore confirm the importance of permafrost carbon release as a positive feedback, and the need to include it accurately in Earth system models, but they do not support considering it to exhibit threshold behaviour.

Clathrates: Some economic assessments continue to emphasise the potential damage from very strong and rapid methane hydrate release (Hope and Schaefer, 2016), although AR5 did not consider this likely. Recent measurements of methane fluxes from the Siberian Shelf Seas (Thornton *et al.*, 2016) are much lower than those inferred previously (Shakhova *et al.*, 2014). A range of other studies have suggested a much smaller influence of clathrate release on the Arctic atmosphere than had been suggested (Berchet *et al.*, 2016; Myhre *et al.*, 2016). New modelling work confirms (Kretschmer *et al.*, 2015) that the Arctic is the region where methane release from clathrates is likely to be most important in the next century, but still estimates methane release to the water column to be negligible compared to anthropogenic releases to the atmosphere. A recent review (Ruppel and Kessler, 2017) emphasises that there remains little evidence that clathrate methane is reaching the atmosphere at present. Although methane that is oxidised in the water column will not reach the atmosphere, it will have the effect of further lowering the pH of the ocean (Boudreau *et al.*, 2015). A recent modelling study joined earlier papers in assigning a relatively limited role to dissociation of methane hydrates as a climate feedback (Mestdagh *et al.*, 2017). Methane concentrations are rising globally, raising interesting questions (see section on methane) about what the cause is (Nisbet *et al.*, 2016; Rigby *et al.*, 2017; Schaefer *et al.*, 2016; Turner *et al.*, 2017). Finally new measurements of the ^{14}C content of methane across the warming out of the last glacial period (Petrenko *et al.*, 2017) show that the release of old carbon reservoirs (including methane hydrates) played only a small role in the methane concentration increase that occurred then.

Amazon rainforest: The processes acting on tropical rainforests are very complex and a recent review still emphasised the possibility of a climate threshold (Nobre *et al.*, 2016). There has been a large body of work looking in theory and practice at bistability between forest and savanna (e.g. Staal *et al.*, 2016). Much of this work accepts that bistability can occur, but raises questions about the spatial scale over which coherent change is likely. There has been considerable progress in investigating future changes in terrestrial carbon, as discussed in the following section. However, a study using multiple models and a range of climate and land use scenarios (Zhang *et al.*, 2015) supports the more cautious IPCC position. Recent work using a detailed ecosystem model (Levine *et al.*, 2016) supports the possibility of a significant but heterogeneous transition in biomass type dependent on the length of the dry season, but in a continuous rather than “tipping point” manner. Resilience may be underestimated if plant trait diversity is not included in models (Sakschewski *et al.*, 2016).

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Is the land taking up carbon dioxide because of faster plant growth?

SUPPLEMENTARY INFORMATION

The IPCC quotes used in this section comes from IPCC AR5, WG1, Chapter 6, Box 6.3.

Global observations

Several lines of evidence point to an increase in primary production over the industrial period. The amplitude of the seasonal cycle of atmospheric CO_2 has been increasing over the last half century, by almost 60% at high northern latitudes (Graven *et al*. 2013). Over the satellite period, observations of vegetation cover show greening (leaf cover) in most forested regions (Zhu *et al*. 2016). Both signals are direct evidence of land taking up more carbon, primarily as a result of CO_2 fertilisation.

The increase in the seasonal amplitude of atmospheric CO_2 is due to increased seasonality in the terrestrial ecosystems in the mid- and high latitudes of the Northern Hemisphere, due to enhanced gross primary production (GPP) in spring and summer, but also enhanced ecosystem respiration across the around the year. The GPP increase is partly due to greening (more/larger leaves are available to harvest sunlight for photosynthesis), but is also due to increasing light use efficiency (LUE: the rate of CO_2 uptake per unit vegetation cover) (Thomas *et al*. 2016).

The observed greening is largely attributed to the fertilisation effect, CO_2-induced increase in GPP leading to more carbon allocation to leaves. The effect of climate, such as warming at high latitudes, is also believed to contribute to the observed greening (Zhu *et al*. 2016). Greening is also found in semi-arid regions, consistent with GPP increase as a result of enhanced water use efficiency (Donohue *et al*. 2013, Ukkola *et al*. 2016).

Global models

The two CMIP5 ESMs that included a coupled land carbon and nitrogen (C-N) cycle greatly underestimated the increase in amplitude of the seasonal cycle of atmospheric CO_2 (Graven *et al*. 2013, Wenzel *et al*. 2016), as well as the interannual variability of atmospheric CO_2 and its sensitivity to climate anomalies (Wenzel *et al*. 2014). Many other land ecosystem models now include a coupled C-N cycle. Most of them produce a stronger CO_2 fertilization effect than the single land model represented in CMIP5. But in general they do so by the wrong mechanism: that is, they allow a shift towards less N-rich tissues (not a major effect as seen experimentally), but they do not allow plants to ‘mine’ additional N from soil (as they have been observed to do in experiments). See Zaehle *et al*. (2014) for a detailed analysis.

We note that Smith *et al*. (2016) claimed that NPP has not increased as much as models calculate. However, this claim is incorrect, being based primarily on evidence from the satellite-derived MODIS NPP 'product' – a model, not data – which considers CO_2 effects only in so far as they are caused by greening, and excludes any possible effect of CO_2 on LUE (De Kauwe *et al*. 2016). Campbell *et al*. (2017), using observed carbonyl sulfide concentration over the 20th century, provided further, independent evidence that GPP increased by about 30% during that period. This increase is towards the upper end of estimates from models.

Limits to CO_2 fertilization

Most plants form symbiotic associations with fungi, which facilitate the supply of soil nutrients to the plants in exchange for a carbon subsidy. Based on the available experimental data, it appears that plants forming symbiotic associations with ectomycorrhizal (ECM) fungi follow one pattern of response to increased CO_2 – with N uptake increasing to meet the increased plant demand; while plants forming associations with arbuscular mycorrhizal (AM) fungi follow a different pattern – with a positive CO_2 response seen only when N availability is high (Terrer *et al*. 2016, 2017). This pattern is consistent with the fact that ECM, and not AM, fungi produce extracellular enzymes that can break down organic forms of N. Nonetheless, further experiments are needed to confirm the generality of this pattern (Norby *et al*. 2017). Knowledge of plant responses to elevated CO_2 in phosphorus-limited ecosystems is very limited (Ellsworth *et al*. 2017) and there are still no experimental data on the CO_2 response of mature tropical forests.

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How do increasing carbon dioxide concentrations impact ocean life and fisheries?

SUPPLEMENTARY INFORMATION

The IPCC quotes used in this section come from IPCC AR5, Synthesis Report, Summary for policymakers, 2.3 and IPCC AR5, Synthesis Report, Future Changes 2.3.1 respectively.

Ocean warming

Ocean warming is now occurring in all ocean basins, at all water depths (Cheng *et al*., 2017; WMO, 2017), although greatest in near-surface waters. The 2014 – 2016 El Niño accentuated that trend, causing extensive coral bleaching, when coral polyps expel algae that live inside their tissues turning the corals white, in the Indian and Pacific Oceans. In the Great Barrier Reef, the additional thermal stress from El Niño proved the most severe that has been recorded, affecting around 90% of individual reefs. It was the third global-scale mass bleaching event since such impacts were first documented in the 1980s (Hughes *et al*., 2017). If current rates of emissions continue most tropical coral reefs ecosystems are unlikely to survive the increase in sea surface temperature in the same form, with greatly diminished goods and services (Hoegh-Guldberg *et al*., 2017). Extensive kelp forests have become degraded with rocky reefs overgrown by turf algae (Filbee-Dexter *et al*., 2016) in warmer waters; these also provide food and habitat for many diverse organisms as well as dampen wave energy around coasts. Warming is shifting the distribution of many species as they seek their optimal temperature. As higher water temperatures alter the habitats on which they depend novel interactions between previously separated species have the potential to transform entire ecological communities (Vergés *et al*., 2016). This may shift fishing grounds towards higher latitudes (Cheung *et al*., 2015) with decreases in fisheries production in some low-latitude countries (Monllor-Hurtado *et al*., 2017). Locations for mariculture are expected to be altered, also resulting in changes in yield patterns among countries (Golden *et al*., 2016).

Ocean deoxygenation

An anthropogenic deoxygenation (decreasing oxygen concentration) signal can already be observed (Long *et al*., 2016) and recent analysis of extensive observational data sets indicate that deoxygenation is occurring more rapidly than previously projected by models (Ito *et al*., 2017) with a whole-ocean 2% loss of oxygen since 1960 (Schmidtko *et al*., 2017), although with high spatial variability. The biogeochemical feedbacks, such as phosphorus-oxygen interactions, remain unclear (Niemeyer *et al*., 2017). Deoxygenation is important in both coastal and open water environments (Levin and Breitburg 2015). In some regions the “metabolic index” (an index based on the oxygen requirement for different animals) is projected to decline greatly (by c. 50% for unabated emissions; Deutsch *et al*. 2015). Future intensification of deoxygenation is expected to have negative impacts on seafloor community structure, on diversity and on fisheries based on more aerobically-active and muscular species of fish, for example tuna (Gallo and Levin 2016).

Ocean acidification

Understanding of ocean acidification and its impacts has significantly improved through greater geographical coverage of integrated chemical and biological observations but uncertainties remain on potential impact particularly in some regions where observations and studies are still lacking. New meta-analyses and assessments have confirmed previous understanding and have also provided additional insights (CBD 2014; Gattuso *et al*., 2015; Nagelkerken and Connell 2015). New studies have demonstrated greater variability in pH in coastal waters and shelf seas than in the open ocean (Chan *et al*., 2017; Ostle *et al*., 2016), biological responses (Hall-Spencer and Allen 2015; Nagelkerken and Munday 2016; Ellis *et al*., 2017) and the complexity of multi-stressor interactions (Calosi *et al*., 2017). Other studies have shown that the carbonate mineral, aragonite, which is vital for shell building, becomes undersaturated or depleted in Arctic waters (Qi *et al*., 2017) and that undersaturation is imminent in the Southern Ocean (Hauri *et al*., 2016); tropical coral reefs are growing slower (Albright *et al*., 2016; Silverman *et al*., 2016); US Pacific coast oyster hatcheries are being impacted (Barton *et al*., 2015) and the shells of small marine swimming snails, pteropods, which are key food web organisms are dissolving off the Californian coast and Southern Ocean (Bednaršek *et al*., 2014).

Combined impact from ocean stressors

Studies on the effects of the multiple stressors have indicated that the scale of adverse biodiversity, ecosystem and fisheries impacts will be influenced by many factors, including complex and not fully understood interactions between ocean warming, deoxygenation and acidification (Kroeker *et al*., 2017) and other non-climate stressors (pollution and over-exploitation and destructive fishing practices). Local variability in conditions, species' responses and adaptation potential, predator-prey interactions, migration of species and complex food-web effects could all affect marine biodiversity, habitats, ecosystems and dependent fisheries (Howes *et al*., 2015; Nagelkerken and Connell 2015).

There is increasing evidence that a high emissions scenario will rapidly and significantly alter many ecosystems and food webs through increased warming, acidification and deoxygenation or their combination. These stressors can represent high or very high risk to fin fisheries and shellfish aquaculture in vulnerable regions and very high risk to tropical coral reefs (Gattuso *et al*., 2015). The potential loss of tropical coral reefs would have major consequences for coastal protection, tourism, income, livelihoods and fisheries (CBD 2014; Hoegh-Guldberg *et al*., 2017). A low emissions scenario (consistent with keeping global temperature increase below 2°C) reduces risk considerably but not entirely; for example, risk to coral ecosystems under the low emissions scenario remains high (Gattuso *et al*., 2015).

Analysis of data on dietary nutrition and the decline in fish catch indicates that the nutrition (protein and micronutrient supply) of about 1.4 billion people is at risk because fish make up >20% of their animal-based food by weight (Golden *et al*., 2016). Destructive fishing practices, industrial pollution, coastal development and climate change contribute to the decline. Climate change under a high emissions scenario is projected, by a study of three models, to reduce fish catch by 6% globally and by 30% in tropical regions by 2050 (Cheung *et al*., 2016) where, due to their dependency on wild fish (Golden *et al*., 2016) and their poor current adaptive capacity (Whitney *et al*., 2017), the most vulnerable communities live.

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How will climate change affect food production on land?

SUPPLEMENTARY INFORMATION

The IPCC quotes used in this section comes from IPCC AR5, Synthesis Report, Future Changes, 2.3.1.

Some studies since 2013 suggest the analysis upon which this statement is based may have somewhat overestimated impacts (Challinor *et al*., 2015), and more regional nuance has emerged, with CO_2 fertilization in some regions increasing yields for some crops (Deryng *et al*., 2014) amid the overall global productivity declines. The conclusions are not altered (Zhao *et al*., 2017), but the new studies point strongly to the importance of accounting for land use / cropping intensity (Challinor *et al*., 2015).

Since 2013, there has been more emphasis on examining the impacts of climate change on nutrition, and not only on yield change. Higher-yielding wheat crops adapted to higher temperatures (Asseng *et al*., 2014), or growing under increased atmospheric CO_2 concentrations (Myers *et al*., 2015; Smith *et al*., 2017), may produce grain of poorer nutritional quality under climate change, for example, with zinc and iron deficiency projected to increase with increasing atmospheric CO_2 concentrations (Myers *et al*., 2015; Smith *et al*., 2017).

AR5 considered productivity changes in yields for the three major cereals. Subsequent analyses have examined potential changes in food production as a result of changes in the area suitable for agriculture (Zabel *et al*., 2014), though greater confidence in the conclusions of such studies requires the inclusion of other criteria for crop growth and development, soil nutrient availability and the incorporation of uncertainty and sensitivity analyses.

Since 2013, we can also say more about the impact of climate change on other crops and rangelands (Havlik *et al*., 2015; Chang *et al*., 2015). Impacts on rangelands sometimes do not translate into large impacts on animal production because of the capacity of intensifying livestock through production system transitions, diet supplementation and other means.

The need for transformation and adaptation in agriculture might be large, and different depending on the climate scenario and socio-technical development pathways (Leclere *et al*., 2014). Since 2013, the costs of adaptation for agriculture have been estimated to be 3% of total agricultural production costs in 2045 (145 billion USD; Weindl *et al*., 2015), somewhat higher than reflected in the literature available in 2013, with the same study also suggesting that, since grass yields are less affected by climate change than arable crop yields, production system shifts towards mixed livestock-cropping systems appear to be a cost effective adaptation option.

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What is the influence of climate change on water availability across the globe?

SUPPLEMENTARY INFORMATION

The IPCC quotes used in this section are from IPCC AR5, WGII, Summary for Policymakers, B-2.

Many catchment-scale studies using CMIP5/RCP scenarios have been published since AR5, all providing quantitative projections. A few global-scale assessments have also been published (e.g. Arnell & Lloyd-Hughes, 2014), including several studies which have used the same set of climate scenarios (e.g. Davie *et al.*, 2013, Wada & Bierkens 2014, Hattermann *et al.*, 2017, Gosling *et al.*, 2017). Koirala *et al.* (2014) and Sedlacek & Knutti (2014) analysed directly the runoff as simulated by CMIP5 models.

Studies of precipitation change in the CMIP5 climate models (e.g. Kent *et al.*, 2015; Chadwick *et al.*, 2016) have shown that the earlier general conclusion that 'wet areas get wetter and dry areas get drier' does not always hold at the regional scale or in each season, due to changes in convection and atmospheric circulation.

Several studies (e.g. Vetter *et al.*, 2017) have also shown the effect of using multiple hydrological models to estimate impacts in order to more comprehensively assess uncertainties. The effects of hydrological model uncertainty are generally smaller than the effects of differences between climate scenarios, but can add substantially to the projected range in hydrological changes – particularly for changes in low flows (Vetter *et al.*, 2017). Studies have compared impacts using different classes of model. Hattermann *et al.* (2016) and Gosling *et al.* (2016) compared global and regional hydrological models, showing that whilst the median changes in runoff from large basins in each class of model were similar, the spread in changes was greater with the global models than with the regional models which were tuned to local conditions. An implication is that projections of regional hydrological changes based on global hydrological models overestimate the effect of hydrological model uncertainty.

Davie *et al.* (2013) showed that global hydrological models could give qualitatively different changes to global biome models (which include CO2 impacts on vegetation and typically model changes in vegetation over time) in some regions, because of the differences in the treatment of evaporation between the two types of model. At the more local scale, Cheng *et al.* (2014) and Ukkola *et al.* (2016a) showed that incorporating the effects of CO2 enrichment on plant respiration and plant growth could substantially change the apparent impacts of climate change.
In semi-arid basins in Australia, increasing plant growth due to CO2 fertilisation was projected to lead to a greater reduction in runoff than would be the case if vegetation cover remained constant (Ukkola *et al.*, 2016a).

This was consistent with observations from Australian rivers (Ukkola *et al*., 2016b). In forested Australian catchments, increasing CO2 concentrations had a greater simulated effect on runoff than changes in climate (Cheng *et al*., 2014).

Hydrological models typically use potential evaporation as an input or calculate potential evaporation from meteorological data. Potential evaporation is the demand for evaporation from the atmosphere – and is therefore arguably a theoretical concept rather than a measurable quantity – and is calculated using a range of approaches, some of which are very empirical. All climate models use the theoretically-based Penman-Monteith formula to derive potential evaporation, but most hydrological models – especially at the catchment scale – do not. Different ways of calculating potential evaporation lead to different estimates of the effect of climate change (Milly & Dunne, 2017), although the difference varies between catchments: it was found to be small compared to scenario uncertainty in the Mekong catchment (Thompson *et al*., 2014) but large in the Delaware catchment (Williamson *et al*., 2016), and large in summer across the US (Chang *et al*., 2016).

Several published studies have searched for local and regional trends in observed hydrological behaviour which can be attributed to climate change. However, hydrological variability is strongly influenced by patterns of internal climatic variability (such as the North Atlantic Oscillation or ENSO), records are often short, and river flows are often affected by human interventions in the catchment such as land use change or water management measures. Alkama *et al*. (2013) attempted to assess trends at the global scale in large basins, but records extended only to 2004 (and most often only to 1992). They found no statistically significant change over the period from 1958 to 2004 from a sample of 161 large basins, but some evidence of change from a larger sample of reconstructed river flows: they concluded that 'changes in global runoff are still unclear', and that methodological assumptions strongly influenced conclusions.

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What is the influence of climate change on species extinction?

SUPPLEMENTARY INFORMATION

The IPCC quotes used in this section comes from IPCC AR5, WGII, Summary for Policymakers, B-2.

A meta-analysis of 131 published studies (all of which had assessed multiple species) combined predictions from different modelling approaches. There were minor differences between correlative and mechanistic models which predicted similar extinction rates that were significantly lower than two other, less commonly used methods. Extinction risks were highest for endemic (small range) species, and in South America, Australia, and New Zealand, and did not vary systematically by taxonomic group (Urban *et al*., 2015) . Linking integrated assessment models with biodiversity models showed how emissions, consumption and economic growth affect biodiversity declines. Extinction risk increases for 8 – 23% of the species, depending on assumptions about species responses to climate change (Visconti *et al*., 2015). Species distribution modelling techniques applied across oceans predict an average poleward latitudinal shift of 15.5 and 25.6 km per decade for a low and high emissions climate change scenario, respectively. Local invasion intensity was highest in high latitude regions, while local extinctions were concentrated near the equator (Jones *et al*., 2015). There are large uncertainties associated with all these estimates, in part because of assumptions in the modelling processes but also the intricate and hard-to-predict biological responses. In 2014 IPCC noted that, "*Some aspects leading to uncertainty in the quantitative projections of extinction risks were not taken into account in previous models; as more realistic details are included, it has been shown that the extinction risks may be either under- or overestimated when based on simpler models.* This statement remains true today.

Research is starting to unravel key factors that may cause under- or over-predicted rates of species loss, including where non-climatic range limitations are common (Early *et al*. 2014), examining ecological correlates of climate tolerances (Estrada *et al*., 2015) and observed responses to climate change (McCain *et al*., 2014).

The importance of spatial resolution in modelling has been emphasised including an increasing appreciation of the importance of microclimates for species survival (micro-refugia), buffering some species (for example, plants, invertebrates) from climate changes especially in terrestrial environments (Maclean *et al*., 2015; Suggitt at al., 2015). Models for many large mammals and birds use climate data with a spatial resolution that is a biologically relevant area for population dynamics, but not always for many small mammals, reptiles and amphibians, and is likely to reduce the predictive accuracy (Nadeau *et al*., 2017). Species with narrow latitudinal ranges are more often limited by factors other than climate (Sunday *et al*., 2015).

Empirical and experimental studies are providing new insights into how changes to local climatic conditions actually play out in population dynamics and range shifts, based on cross species studies (Pearson *et al*., 2014), experiments and observations (Bestion *et al*., 2015; Wiens *et al*., 2016).

Many climate change studies have ignored other environmental changes and stressors despite the evidence that the interactions can be significant (Pacifici *et al*., 2015; Chapman *et al*., 2014; Selwood *et al*., 2015) and there is emerging evidence that loss of interaction partners is a key factor affecting the persistence of many species (Brodie *et al*., 2014).

A major caveat in almost all climate change and species extinction studies is weak consideration of adaptive responses. Evolution can occur rapidly, but may be difficult to sustain at high rates in a changing environment. A predictive framework for the genetic architecture underlying adaptive responses is starting to emerge (Bay *et al*., 2017). Persistence can be greatly facilitated by non-genetic adaptive responses (including phenotypic plasticity), and heritable variation in plasticity can further speed up genetic evolution (Chevin *et al*., 2015). Adaptation to a new extreme environment may often occur by a rapid transient increase in plasticity followed by slow genetic assimilation involving decreased plasticity (Lande *et al*., 2015).

Different species respond differently to climate change, and therefore including species interactions with other species (for example herbivory, pathogens, parasites, competitors, predators) into climate change models improves model accuracy, but also increases the diversity of predicted outcomes (Staniczenko *et al*., 2017). Recent models of adaptation in multispecies communities indicate that the potential for rapid evolution is strongly influenced by the extent of competition among species (Kopp *et al*., 2014).

Given the complexity of species responses, there has been an increased emphasis on providing guidance to practitioners on how to assess species at risk from climate change. Sidestepping the extreme and intensive data and information demands of the most effective modelling techniques has become important for decisions being taken now. There are new guidelines (Pacifici *et al*., 2015) but this is still an important area for research (Urban *et al*., 2016).

An earlier concern was that protected areas could be in the wrong places, given that species may need to move to cope with a changing climate. However, better evidence now exists showing that protected areas remain important for conservation even if it is for different species (Thomas *et al*., 2015).

Extended figure legend:

Extinction risks for four scenarios are provided: the current post-industrial temperature rise of 0.8°C [2011 – 2015 average; WMO, 2016], the policy target of 2°C, and representative concentration pathways (RCPs) 6.0 and 8.5. Pre-industrial rise was calculated using standard methods (Warren *et al*., 2011). Results are based on more than 550,000 predictions from 131 multi-species studies of extinction risk from climate change from a meta-analysis in Urban (Urban *et al*., 2015). Error bars indicate 95% credible intervals from the significant regression of predicted extinction risk against preindustrial temperature rise.

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How will aspects of human health be affected by climate change?

SUPPLEMENTARY INFORMATION

The IPCC quotes used in this section comes from IPCC AR5, WGII, Summary for Policymakers, section B-2.

Increasing evidence is accruing of growing population exposure to hazardous levels of heat stress under climate change. For example even if temperature increase is limited to 2 °C above pre-industrial, Karachi (Pakistan) and Kolkata (India) could experience conditions equivalent to their deadly 2015 heatwaves every year. With only 1.5 °C of global warming and midrange population growth, over 350 million more people could be exposed to such levels of heat by 2050 in cities such as Lagos, Nigeria, and Shanghai, China (Matthews *et al.*, 2016).

An analysis of 783 cases of excess human mortality associated with heat from 164 cities in 36 countries showed a global threshold beyond which daily mean surface air temperature and relative humidity increase risk of death. Around 30% of the world's population is currently exposed, for at least 20 days a year, to climatic conditions exceeding this threshold. By 2100, this percentage is projected to increase to c. 48% under RCP 4.5 and c. 74% under RCP 8.5 using population projections from the Shared Socioeconomic Pathways (SSPs) for each RCP (Mora *et al.*, 2017).

A study of ten major cities in the USA estimated the change in temperature-related mortality rates in 2045 – 2055 and 2085 – 2095 compared to 1992 – 2002, under two greenhouse gas emissions scenarios (RCP 4.5 and 8.5). Under the higher-emission scenario, 8 of the 10 metropolitan areas are projected to experience a net increase in annual temperature-related deaths per million people by 2086 – 2095, with the net increase in Los Angeles of 627 (95% empirical CI: 239-1018) deaths per million (Weinberger *et al.*, 2017). In seven major Korean cities significantly positive associations between ambient temperature during 1 month before death or 2 weeks before death and total infant mortality or mortality from SIDS were found. The overall hazard ratio of infant mortality from total deaths and SIDS for a 1 °C increase during 1 month before death was 1.52 (95% CI, 1.46–1.57) and 1.50 (95% CI, 1.35–1.66), respectively (Son *et al.*, 2017).

Data from 451 locations in 23 countries across nine regions of the world, including over 85 m deaths in the period 1984 – 2015 were used to estimate temperature-mortality relationships. Excess mortality for cold, heat, and their net change was projected in 1990 – 2099 under climate change scenarios determined by varying trajectories of greenhouse gas emissions, assuming no adaptation or population changes. On average, there was a net increase in temperature-related excess mortality under high-emission scenarios, although in temperate areas such as North Europe, and East Asia, the large projected decrease in cold-related excess would result in a null or marginally negative net impact. Conversely, warmer regions, such as the central and southern parts of Europe, and especially South-East Asia (no data were available for Africa), would experience a sharp surge in heat-related impacts and extremely large net mortality increases. Most of the health effects directly due to temperature increase could be avoided by mitigation strategies to limit emissions and further warming of the planet.

Several modelling studies since 2013 have confirmed that dengue is projected to expand, particularly at the edge of current distribution ranges (Ebi and Nolan 2016). One study suggests that the population exposure to the main mosquito vector (which also transmits Zika) would increase by 8 – 12% due to climate change alone and by 127 – 134% when climate change is combined with a development pathway associated with high population growth (Monaghan *et al*. 2016).

A study linking a detailed agricultural modelling framework to a comparative risk assessment of changes in fruit and vegetable consumption, red meat consumption, and bodyweight for deaths from a number of causes, projected that by 2050, climate change could lead to per-person reductions of 3·2% (SD 0·4%) in global food availability, 4·0% (0·7%) in fruit and vegetable consumption, and 0·7% (0·1%) in red meat consumption. These declines are projected to lead to an estimated net 529 000 extra climate change-related deaths worldwide (95% CI 314 000–736 000), mainly in south and east Asia. (Springmann *et al*., 2016).

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