

Chapter 2: Land-Climate Interactions

Coordinating Lead Authors: Gensuo Jia (China), Elena Shevliakova (The United States of America)

Lead Authors: Paulo Artaxo (Brazil), Nathalie De Noblet-Ducoudré (France), Richard Houghton (The United States of America), Joanna House (United Kingdom), Kaoru Kitajima (Japan), Christopher Lennard (South Africa), Alexander Popp (Germany), Andrey Sirin (The Russian Federation), Raman Sukumar (India), Louis Verchot (Colombia/The United States of America)

Contributing Authors: William Anderegg (The United States of America), Edward Armstrong (United Kingdom), Ana Bastos (Portugal/Germany), Terje Koren Bernsten (Norway), Peng Cai (China), Katherine Calvin (The United States of America), Francesco Cherubini (Italy), Sarah Connors (France/United Kingdom), Annette Cowie (Australia), Edouard Davin (Switzerland/ France), Cecile De Klein (New Zealand), Giacomo Grassi (Italy/EU), Rafiq Hamdi (Belgium), Florian Humpenöder (Germany), David Kanter (The United States of America), Gerhard Krinner (France), Sonali McDermid (India/The United States of America), Devaraju Narayanappa (India/France), Josep Peñuelas (Spain), Prajal Pradhan (Nepal), Benjamin Quesada (Colombia), Stephanie Roe (The Philippines/The United States of America), Robert A. Rohde (The United States of America), Martijn Slot (Panama), Rolf Sommer (Germany), Moa Sporre (Norway), Benjamin Sulman (The United States of America), Alasdair Sykes (United Kingdom), Phil Williamson (United Kingdom), Yuyu Zhou (China/The United States of America)

Review Editors: Pierre Bernier (Canada), Jhan Carlo Espinoza (Peru), Sergey Semenov (The Russian Federation)

Chapter Scientist: Xiyang Xu (China)

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1	Table of Contents	
2	Chapter 2: Land-Climate Interactions	1
3	Executive Summary.....	3
4	2.1 Introduction: Land – climate interactions.....	9
5	2.1.1 Recap of previous IPCC and other relevant reports as baselines.....	9
6	2.1.2 Introduction to the chapter structure	11
7	Box 2.1: Processes underlying land-climate interactions	12
8	2.2 The effect of climate variability and change on land	14
9	2.2.1 Overview of climate impacts on land	14
10	2.2.2 Climate driven changes in aridity	16
11	2.2.3 The influence of climate change on food security	17
12	2.2.4 Climate-driven changes in terrestrial ecosystems	17
13	2.2.5 Climate extremes and their impact on land functioning	19
14	Cross-Chapter Box 3: Fire and Climate Change	25
15	2.3 Greenhouse gas fluxes between land and atmosphere.....	28
16	2.3.1 Carbon Dioxide.....	29
17	2.3.2 Methane	36
18	2.3.3 Nitrous Oxide.....	40
19	Box 2.2: Methodologies for estimating national to global scale anthropogenic land carbon fluxes..	43
20	2.4 Emissions and impacts of short-lived climate forcers (SLCF) from land	46
21	2.4.1 Mineral dust	47
22	2.4.2 Carbonaceous Aerosols.....	49
23	2.4.3 Biogenic Volatile Organic Compounds (BVOCs).....	51
24	2.5 Land impacts on climate and weather through biophysical and GHGs effects	53
25	2.5.1 Impacts of historical and future anthropogenic land cover changes	54
26	2.5.2 Impacts of specific land use changes	60
27	2.5.3 Amplifying / dampening climate changes via land responses	68
28	2.5.4 Non-local and downwind effects resulting from changes in land cover	72
29	Cross-Chapte Box 4: Climate Change and Urbanisation	73
30	2.6 Climate consequences of response options	77
31	2.6.1 Climate impacts of individual response options	77
32	2.6.2 Integrated pathways for climate change mitigation	85
33	2.6.3 The contribution of response options to the Paris Agreement	91
34	2.7 Plant and soil processes underlying land-climate interactions	94
35	2.7.1 Temperature responses of plant and ecosystem production.....	94
36	2.7.2 Water transport through soil-plant-atmosphere continuum and drought mortality.....	95
37	2.7.3 Soil microbial effects on soil nutrient dynamics and plant responses to elevated CO ₂	96
38	2.7.4 Vertical distribution of soil organic carbon	97
39	2.7.5 Soil carbon responses to warming and changes in soil moisture	97
40	2.7.6 Soil carbon responses to changes in organic-matter inputs by plants	98
41	References	101
42	Appendix	178
43		
44		
45		

1 **Executive Summary**

2
3 Land and climate interact in complex ways through changes in forcing and multiple biophysical and
4 biogeochemical feedbacks across different spatial and temporal scales. This chapter assesses climate impacts
5 on land and land impacts on climate, the human contributions to these changes, as well as land-based
6 adaptation and mitigation response options to combat projected climate changes.
7

8 ***Implications of climate change, variability, and extremes for land systems***

9
10 **It is certain that globally averaged land surface air temperature (LSAT) has risen faster than the**
11 **global mean surface temperature (i.e., combined LSAT and sea surface temperature) from**
12 **preindustrial (1850–1900) to present day (1999–2018). According to the single longest and most**
13 **extensive dataset, the LSAT increase between the preindustrial period and present day was 1.52°C**
14 **(the *very likely* range of 1.39°C to 1.66°C). For the 1880–2018 period, when four independently**
15 **produced datasets exist, the LSAT increase was 1.41°C (1.31°C–1.51°C), where the range represents**
16 **the spread in the datasets' median estimates.** Analyses of paleo records, historical observations, model
17 simulations, and underlying physical principles are all in agreement that LSATs are increasing at a higher
18 rate than SST as a result of differences in evaporation, land-climate feedbacks, and changes in the aerosol
19 forcing over land (*very high confidence*). For the 2000–2016 period, the land-to-ocean warming ratio (about
20 1.6) is in close agreement between different observational records and the CMIP5 climate model simulations
21 (the *likely* range of 1.54 to 1.81). {2.2.1}

22
23 **Anthropogenic warming has resulted in shifts of climate zones, primarily as an increase in dry**
24 **climates and decrease of polar climates (*high confidence*). Ongoing warming is projected to result in**
25 **new, hot climates in tropical regions and to shift climate zones poleward in the mid- to high latitudes**
26 **and upward in regions of higher elevation (*high confidence*).** Ecosystems in these regions will become
27 increasingly exposed to temperature and rainfall extremes beyond climate regimes they are currently adapted
28 to (*high confidence*), which can alter their structure, composition and functioning. Additionally, high-latitude
29 warming is projected to accelerate permafrost thawing and increase disturbance in boreal forests through
30 abiotic (e.g., drought, fire) and biotic (e.g., pests, disease) agents (*high confidence*). {2.2.1, 2.2.2, 2.5.3}

31
32 **Globally, greening trends (trends of increased photosynthetic activity in vegetation) have increased**
33 **over the last 2-3 decades by 22–33%, particularly over China, India, many parts of Europe, central**
34 **North America, southeast Brazil and southeast Australia (*high confidence*).** This results from a
35 combination of direct (i.e., land use and management, forest conservation and expansion) and indirect factors
36 (i.e., CO₂ fertilisation, extended growing season, global warming, nitrogen deposition, increase of diffuse
37 radiation) linked to human activities (*high confidence*). Browning trends (trends of decreasing photosynthetic
38 activity) are projected in many regions where increases in drought and heat waves are projected in a warmer
39 climate. There is *low confidence* in the projections of global greening and browning trends. {2.2.4, Cross-
40 Chapter Box 4: Climate change and urbanisation, in this chapter}

41
42 **The frequency and intensity of some extreme weather and climate events have increased as a**
43 **consequence of global warming and will continue to increase under medium and high emission**
44 **scenarios (*high confidence*).** Recent heat-related events, e.g., heat waves, have been made more frequent or
45 intense due to anthropogenic greenhouse gas emissions in most land regions and the frequency and intensity
46 of drought has increased in Amazonia, north-eastern Brazil, the Mediterranean, Patagonia, most of Africa
47 and north-eastern China (*medium confidence*). Heat waves are projected to increase in frequency, intensity
48 and duration in most parts of the world (*high confidence*) and drought frequency and intensity is projected to
49 increase in some regions that are already drought prone, predominantly in the Mediterranean, central Europe,
50 the southern Amazon and southern Africa (*medium confidence*). These changes will impact ecosystems, food
51 security and land processes including greenhouse gas (GHG) fluxes (*high confidence*). {2.2.5}

1
2 **Climate change is playing an increasing role in determining wildfire regimes along-side human**
3 **activity (*medium confidence*), with future climate variability expected to enhance the risk and severity**
4 **of wildfires in many biomes such as tropical rainforests (*high confidence*).** Fire weather seasons have
5 lengthened globally between 1979 and 2013 (*low confidence*). Global land area burned has declined in recent
6 decades, mainly due to less burning in grasslands and savannahs (*high confidence*). While drought remains
7 the dominant driver of fire emissions, there has recently been increased fire activity in some tropical and
8 temperate regions during normal to wetter than average years due to warmer temperatures that increase
9 vegetation flammability (*medium confidence*). The boreal zone is also experiencing larger and more frequent
10 fires, and this may increase under a warmer climate (*medium confidence*). {Cross-Chapter Box 4: Climate
11 change and urbanisation, in this chapter}

12 ***Terrestrial greenhouse gas fluxes on unmanaged and managed lands***

13 **Agriculture, Forestry and Other Land Use (AFOLU) is a significant net source of GHG emissions**
14 **(*high confidence*), contributing to about 22% of anthropogenic emissions of carbon dioxide (CO₂),**
15 **methane (CH₄), and nitrous oxide (N₂O) combined as CO₂ equivalents in 2007 to 2016 (*medium***
16 ***confidence*).** AFOLU results in both emissions and removals of CO₂, CH₄, and N₂O to and from the
17 atmosphere (*high confidence*). These fluxes are affected simultaneously by natural and human drivers,
18 making it difficult to separate natural from anthropogenic fluxes (*very high confidence*). {2.3}

19 **The total net land-atmosphere flux of CO₂ on both managed and unmanaged lands very likely**
20 **provided a global net removal from 2008 to 2017 according to models, (-6.2 ± 3.7 GtCO₂ yr⁻¹, *medium***
21 ***confidence*).** This net removal is comprised of two major components: i) modelled net anthropogenic
22 emissions from AFOLU are *likely* 5.5 ± 2.6 GtCO₂ yr⁻¹ driven by land cover change, including deforestation
23 and afforestation/reforestation, and wood harvesting (accounting for about 13% of total net anthropogenic
24 emissions of CO₂) (*medium confidence*); and ii) modelled net removals due to non-anthropogenic processes
25 are *likely* 11.7 ± 2.6 GtCO₂ yr⁻¹ on managed and unmanaged lands, driven by environmental changes such as
26 increasing CO₂, nitrogen deposition, and changes in climate (accounting for a removal of 29% of the CO₂
27 emitted from all anthropogenic activities (fossil fuel, industry and AFOLU) (*medium confidence*). {2.3.1}

28 **The anthropogenic emissions of CO₂ from AFOLU reported in countries' GHG inventories were 0.1 ±**
29 **1.0 GtCO₂ yr⁻¹ globally during 2005 to 2014 (*low confidence*), much lower than emission estimates**
30 **from global models of 5.1 ± 2.6 GtCO₂ yr⁻¹ over the same time period. Reconciling these differences**
31 **can support consistency and transparency in assessing global progress towards meeting modelled**
32 **mitigation pathway such as under the Paris Agreement's global stocktake (*medium confidence*).** This
33 discrepancy is consistent with understanding of the different approaches used to defining anthropogenic
34 fluxes. Inventories consider larger areas of forested lands as managed than models do, and report all fluxes
35 on managed lands as anthropogenic, including a large net sink due to the indirect effects of changing
36 environmental conditions (e.g., climate change, and change in atmospheric CO₂ and N). In contrast, the
37 models assign part of this indirect forest sink to the non-anthropogenic sink on unmanaged lands. {2.3.1}

38 **The gross emissions from AFOLU (one third of total global emissions) are more indicative of**
39 **mitigation potential of reduced deforestation than the global net emissions (13% of total global**
40 **emissions), which include compensating deforestation and afforestation fluxes (*high confidence*).** The
41 net flux of CO₂ from AFOLU is composed of two opposing gross fluxes: gross emissions (20 GtCO₂ yr⁻¹)
42 from deforestation, cultivation of soils, and oxidation of wood products; and gross removals (-14 GtCO₂ yr⁻¹)
43 largely from forest growth following wood harvest and agricultural abandonment (*medium confidence*).
44 {2.3.1}

45 **Land is a net source of CH₄, accounting for 61% of anthropogenic CH₄ emissions for the 2005–2015**

1 **period (*medium confidence*)**. The pause in the rise of atmospheric CH₄ concentrations between 2000 and
2 2006 and the subsequent renewed increase appear to be partially associated with land use and land use
3 change. The recent depletion trend of the ¹³C isotope in the atmosphere indicates that higher biogenic sources
4 explain part of the current CH₄ increase and that biogenic sources make up a larger proportion of the source
5 mix than they did before 2000 (*high confidence*). In agreement with the findings of AR5, tropical wetlands
6 and peatlands continue to be important drivers of inter-annual variability and current CH₄ concentration
7 increases (*medium evidence, high agreement*). Ruminants and the expansion of rice cultivation are also
8 important contributors to the current trend (*medium evidence, high agreement*). There is significant and
9 ongoing accumulation of CH₄ in the atmosphere (*very high confidence*). {2.3.2}

10
11 **AFOLU is the main anthropogenic sources of N₂O primarily due to nitrogen (N) application to soils**
12 (*high confidence*). In croplands, the main driver of N₂O emissions is a lack of synchronisation between crop
13 N demand and soil N supply, with approximately 50% of the N applied to agricultural land not taken up by
14 the crop. Cropland soils emit over 3 Mt N₂O-N yr⁻¹ (*medium confidence*). Because the response of N₂O
15 emissions to fertiliser application rates is non-linear, in regions of the World where low N application rates
16 dominate, such as sub-Saharan Africa and parts of Eastern Europe, increases in N fertiliser use would
17 generate relatively small increases in agricultural N₂O emissions. Decreases in application rates in regions
18 where application rates are high and exceed crop demand for parts of the growing season will have very
19 large effects on emissions reductions (*medium evidence, high agreement*). {2.3.3}

20
21 **While managed pastures make up only one-quarter of grazing lands, they contributed more than**
22 **three-quarters of N₂O emissions from grazing lands between 1961 and 2014 with rapid recent**
23 **increases of N inputs resulting in disproportionate growth in emissions from these lands (*medium***
24 ***confidence*)**. Grazing lands (pastures and rangelands) are responsible for more than one-third of total
25 anthropogenic N₂O emissions or more than one-half of agricultural emissions (*high confidence*). Emissions
26 are largely from North America, Europe, East Asia, and South Asia, but hotspots are shifting from Europe to
27 southern Asia (*medium confidence*). {2.3.3}

28
29 **Increased emissions from vegetation and soils due to climate change in the future are expected to**
30 **counteract potential sinks due to CO₂ fertilisation (*low confidence*)**. Responses of vegetation and soil
31 organic carbon (SOC) to rising atmospheric CO₂ concentration and climate change are not well constrained
32 by observations (*medium confidence*). Nutrient (e.g., nitrogen, phosphorus) availability can limit future plant
33 growth and carbon storage under rising CO₂ (*high confidence*). However, new evidence suggests that
34 ecosystem adaptation through plant-microbe symbioses could alleviate some nitrogen limitation (*medium*
35 *evidence, high agreement*). Warming of soils and increased litter inputs will accelerate carbon losses through
36 microbial respiration (*high confidence*). Thawing of high-latitude/altitude permafrost will increase rates of
37 SOC loss and change the balance between CO₂ and CH₄ emissions (*medium confidence*). The balance
38 between increased respiration in warmer climates and carbon uptake from enhanced plant growth is a key
39 uncertainty for the size of the future land carbon sink (*medium confidence*). {2.3.1, 2.7.2, Box 2.3}

40 41 ***Biophysical and biogeochemical land forcing and feedbacks to the climate system***

42
43 **Changes in land conditions from human use or climate change in turn affect regional and global**
44 **climate (*high confidence*)**. On the global scale, this is driven by changes in emissions or removals of CO₂,
45 CH₄, and N₂O by land (biogeochemical effects) and by changes in the surface albedo (*very high confidence*).
46 Any local land changes that redistribute energy and water vapour between the land and the atmosphere
47 influence regional climate (biophysical effects; *high confidence*). However, there is *no confidence* in whether
48 such biophysical effects influence global climate. {2.1, 2.3, 2.5.1, 2.5.2}

49
50 **Changes in land conditions modulate the likelihood, intensity and duration of many extreme events**
51 **including heat waves (*high confidence*) and heavy precipitation events (*medium confidence*)**. Dry soil

1 conditions favour or strengthen summer heat wave conditions through reduced evapotranspiration and
2 increased sensible heat. By contrast wet soil conditions, for example from irrigation, or crop management
3 practices that maintain a cover crop all year round, can dampen extreme warm events through increased
4 evapotranspiration and reduced sensible heat. Droughts can be intensified by poor land management.
5 Urbanisation increases extreme rainfall events over or downwind of cities (*medium confidence*). {2.5.1,
6 2.5.2, 2.5.3}

7
8 **Historical changes in anthropogenic land cover have resulted in a mean annual global warming of**
9 **surface air from biogeochemical effects (*very high confidence*), dampened by a cooling from**
10 **biophysical effects (*medium confidence*).** Biogeochemical warming results from increased emissions of
11 GHGs by land, with model-based estimates of $+0.20\pm 0.05^{\circ}\text{C}$ (global climate models) and $+0.24\pm 0.12^{\circ}\text{C}$
12 (dynamic global vegetation models, DGVMs) as well as an observation-based estimate of $+0.25\pm 0.10^{\circ}\text{C}$. A
13 net biophysical cooling of $-0.10\pm 0.14^{\circ}\text{C}$ has been derived from global climate models in response to the
14 increased surface albedo and decreased turbulent heat fluxes, but it is smaller than the warming effect from
15 land-based emissions. However when both biogeochemical and biophysical effects are accounted for within
16 the same global climate model, the models do not agree on the sign of the net change in mean annual surface
17 air temperature. {2.3, 2.5.1, Box 2.1}

18
19 **The future projected changes in anthropogenic land cover that have been examined for AR5 would**
20 **result in a biogeochemical warming and a biophysical cooling whose magnitudes depend on the**
21 **scenario (*high confidence*).** Biogeochemical warming has been projected for RCP8.5 by both global climate
22 models ($+0.20\pm 0.15^{\circ}\text{C}$) and DGVMs ($+0.28\pm 0.11^{\circ}\text{C}$) (*high confidence*). A global biophysical cooling of
23 $0.10\pm 0.14^{\circ}\text{C}$ is estimated from global climate models, and projected to dampen the land-based warming (*low*
24 *confidence*). For RCP4.5 the biogeochemical warming estimated from global climate models ($+0.12\pm 0.17^{\circ}\text{C}$)
25 is stronger than the warming estimated by DGVMs ($+0.01\pm 0.04^{\circ}\text{C}$) but based on *limited evidence*, as is the
26 biophysical cooling ($-0.10\pm 0.21^{\circ}\text{C}$). {2.5.2}

27
28 **Regional climate change can be dampened or enhanced by changes in local land cover and land use**
29 **(*high confidence*) but this depends on the location and the season (*high confidence*).** In boreal regions,
30 for example, where projected climate change will migrate treeline northward, increase the growing season
31 length and thaw permafrost, regional winter warming will be enhanced by decreased surface albedo and
32 snow, whereas warming will be dampened during the growing season due to larger evapotranspiration (*high*
33 *confidence*). In the tropics, wherever climate change will increase rainfall, vegetation growth and associated
34 increase in evapotranspiration will result in a dampening effect on regional warming (*medium confidence*).
35 {2.5.2, 2.5.3}

36
37 **According to model-based studies, changes in local land cover or available water from irrigation**
38 **affect climate in regions as far as few hundreds of kilometres downwind (*high confidence*).** The local
39 redistribution of water and energy following the changes on land affect the horizontal and vertical gradients
40 of temperature, pressure and moisture, thus alter regional winds and consequently moisture and temperature
41 advection and convection, and this affects precipitation. {2.5.2, 2.5.4, Cross-Chapter Box 4: Climate
42 Change and Urbanisation}

43
44 **Future increases in both climate change and urbanisation will enhance warming in cities and their**
45 **surroundings (urban heat island), especially during heat waves (*high confidence*).** Urban and peri-urban
46 agriculture, and more generally urban greening, can contribute to mitigation (*medium confidence*) as well as
47 to adaptation (*high confidence*), with co-benefits for food security and reduced soil-water-air pollution.
48 {Cross-Chapter Box 4: Climate Change and Urbanisation}

49
50 **Regional climate is strongly affected by natural land aerosols (*medium confidence*) (e.g., mineral dust,**
51 **black, brown and organic carbon), but there is *low confidence* in historical trends, interannual and**

1 **decadal variability, and future changes.** Forest cover affects climate through emissions of biogenic
2 volatile organic compounds (BVOC) and aerosols (*low confidence*). The decrease in the emissions of
3 BVOC resulting from the historical conversion of forests to cropland has resulted in a positive radiative
4 forcing through direct and indirect aerosol effects, a negative radiative forcing through the reduction in the
5 atmospheric lifetime of methane and it has contributed to increased ozone concentrations in different
6 regions (*low confidence*). {2.4, 2.5}

7
8 ***Consequences for the climate system of land-based adaptation and mitigation options, including carbon
9 dioxide removal (negative emissions)***

10
11 **About one quarter of the 2030 mitigation pledged by countries in their initial Nationally Determined
12 Contributions (NDCs) under the Paris Agreement is expected to come from land-based mitigation
13 options (*medium confidence*).** Most of the Nationally Determined Contributions (NDCs) submitted by
14 countries include land-based mitigation, although many lack details. Several refer explicitly to reduced
15 deforestation and forest sinks, while a few include soil carbon sequestration, agricultural management and
16 bioenergy. Full implementation of NDCs (submitted by February 2016) is expected to result in net
17 removals of 0.4–1.3 GtCO₂ y⁻¹ in 2030 compared to the net flux in 2010, where the range represents low to
18 high mitigation ambition in pledges, not uncertainty in estimates (*medium confidence*). {2.6.3}

19
20 **Several mitigation response options have technical potential for >3 GtCO₂-eq yr⁻¹ by 2050 through
21 reduced emissions and Carbon Dioxide Removal (CDR) (*high confidence*), some of which compete
22 for land and other resources, while others may reduce the demand for land (*high confidence*).**
23 Estimates of the technical potential of individual response options are not necessarily additive. The largest
24 potential for reducing AFOLU emissions are through reduced deforestation and forest degradation (0.4–5.8
25 GtCO₂-eq yr⁻¹) (*high confidence*), a shift towards plant-based diets (0.7–8.0 GtCO₂-eq yr⁻¹) (*high
26 confidence*) and reduced food and agricultural waste (0.8–4.5 CO₂-eq yr⁻¹) (*high confidence*). Agriculture
27 measures combined could mitigate 0.3–3.4 GtCO₂-eq yr⁻¹ (*medium confidence*). The options with largest
28 potential for CDR are afforestation/reforestation (0.5–10.1 CO₂-eq yr⁻¹) (*medium confidence*), soil carbon
29 sequestration in croplands and grasslands (0.4–8.6 CO₂-eq yr⁻¹) (*high confidence*) and Bioenergy with
30 Carbon Capture and Storage (BECCS) (0.4–11.3 CO₂-eq yr⁻¹) (*medium confidence*). While some estimates
31 include sustainability and cost considerations, most do not include socio-economic barriers, the impacts of
32 future climate change or non-GHG climate forcings. {2.6.1}

33
34 **Response options intended to mitigate global warming will also affect the climate locally and
35 regionally through biophysical effects (*high confidence*).** Expansion of forest area, for example, typically
36 removes CO₂ from the atmosphere and thus dampens global warming (biogeochemical effect, *high
37 confidence*), but the biophysical effects can dampen or enhance regional warming depending on location,
38 season and time of day. During the growing season, afforestation generally brings cooler days from
39 increased evapotranspiration, and warmer nights (*high confidence*). During the dormant season, forests are
40 warmer than any other land cover, especially in snow-covered areas where forest cover reduces albedo
41 (*high confidence*). At the global level, the temperature effects of boreal afforestation/reforestation run
42 counter to GHG effects, while in the tropics they enhance GHG effects. In addition, trees locally dampen
43 the amplitude of heat extremes (*medium confidence*). {2.5.2, 2.5.4, 2.7, Cross-Chapter Box 4: Climate
44 Change and Urbanisation}

45
46 **Mitigation response options related to land use are a key element of most modelled scenarios that
47 provide strong mitigation, alongside emissions reduction in other sectors (*high confidence*).** More
48 stringent climate targets rely more heavily on land-based mitigation options, in particular, CDR
49 (*high confidence*). Across a range of scenarios in 2100, CDR is delivered by both afforestation (median
50 values of -1.3, -1.7 and -2.4 GtCO₂yr⁻¹ for scenarios RCP4.5, RCP2.6 and RCP1.9 respectively) and
51 bioenergy with carbon capture and storage (BECCS) (-6.5, -11 and -14.9 GtCO₂ yr⁻¹). Emissions of CH₄

1 and N₂O are reduced through improved agricultural and livestock management as well as dietary shifts
2 away from emission-intensive livestock products by 133.2, 108.4 and 73.5 MtCH₄yr⁻¹; and 7.4, 6.1 and 4.5
3 MtN₂O yr⁻¹ for the same set of scenarios in 2100 (*high confidence*). High levels of bioenergy crop
4 production can result in increased N₂O emissions due to fertiliser use. The Integrated Assessment Models
5 that produce these scenarios mostly neglect the biophysical effects of land-use on global and regional
6 warming. {2.5, 2.6.2}

7
8 **Large-scale implementation of mitigation response options that limit warming to 1.5 or 2°C would**
9 **require conversion of large areas of land for afforestation/reforestation and bioenergy crops, which**
10 **could lead to short-term carbon losses (*high confidence*).** The change of global forest area in mitigation
11 pathways ranges from about -0.2 to +7.2 Mkm² between 2010 and 2100 (median values across a range of
12 models and scenarios: RCP4.5, RCP2.6, RCP1.9), and the land demand for bioenergy crops ranges from
13 about 3.2–6.6 Mkm² in 2100 (*high confidence*). Large-scale land-based CDR is associated with multiple
14 feasibility and sustainability constraints (Chapters 6, 7). In high carbon lands such as forests and peatlands,
15 the carbon benefits of land protection are greater in the short-term than converting land to bioenergy crops
16 for BECCS, which can take several harvest cycles to ‘pay-back’ the carbon emitted during conversion
17 (carbon-debt), from decades to over a century (*medium confidence*). {2.6.2, Chapters 6, 7}

18
19 **It is possible to achieve climate change targets with low need for land-demanding CDR such as**
20 **BECCS, but such scenarios rely more on rapidly reduced emissions or CDR from forests, agriculture**
21 **and other sectors.** Terrestrial CDR has the technical potential to balance emissions that are difficult to
22 eliminate with current technologies (including food production). Scenarios that achieve climate change
23 targets with less need for terrestrial CDR rely on agricultural demand-side changes (diet change, waste
24 reduction), and changes in agricultural production such as agricultural intensification. Such pathways that
25 minimise land use for bioenergy and BECCS are characterised by rapid and early reduction of GHG
26 emissions in all sectors, as well as earlier CDR in through afforestation. In contrast, delayed mitigation
27 action would increase reliance on land-based CDR (*high confidence*). {2.6.2}

2.1 Introduction: Land – climate interactions

This chapter assesses the literature on two-way interactions between climate and land, with focus on scientific findings published since AR5 and some aspects of the land-climate interactions that were not assessed in previous IPCC reports. Previous IPCC assessments recognised that climate affects land cover and land surface processes, which in turn affect climate. However, previous assessments mostly focused on the contribution of land to global climate change via its role in emitting and absorbing greenhouse gases (GHGs) and short-lived climate forcers (SLCFs), or via implications of changes in surface reflective properties (i.e., albedo) for solar radiation absorbed by the surface. This chapter examines scientific advances in understanding the interactive changes of climate and land, including impacts of climate change, variability and extremes on managed and unmanaged lands. It assesses climate forcing of land changes from direct (e.g., land use change and land management) and indirect (e.g., increasing atmospheric CO₂ concentration and nitrogen deposition) effects at local, regional, and global scale.

2.1.1 Recap of previous IPCC and other relevant reports as baselines

The evidence that land cover matters for the climate system have long been known, especially from early paleoclimate modelling studies and impacts of human-induced deforestation at the margin of deserts (de Noblet et al. 1996; Kageyama et al. 2004). The understanding of how land use activities impact climate has been put forward by the pioneering work of (Charney 1975) who examined the role of overgrazing-induced desertification on the Sahelian climate.

Since then there have been many modelling studies that reported impacts of idealised or simplified land cover changes on weather patterns (e.g., Pielke et al. 2011). The number of studies dealing with such issues has increased significantly over the past 10 years, with more studies that address realistic past or projected land changes. However, very few studies have addressed the impacts of land cover changes on climate as very few land surface models embedded within climate models (whether global or regional), include a representation of land management. Observation-based evidence of land-induced climate impacts emerged even more recently (e.g., Alkama and Cescatti 2016; Bright et al. 2017; Lee et al. 2011; Li et al. 2015; Duveiller et al. 2018; Forzieri et al. 2017) and the literature is therefore limited.

In previous IPCC reports, the interactions between climate change and land were covered separately by three working groups. AR5 WGI assessed the role of land use change in radiative forcing, land-based GHGs source and sink, and water cycle changes that focused on changes of evapotranspiration, snow and ice, runoff, and humidity. AR5 WGII examined impacts of climate change on land, including terrestrial and freshwater ecosystems, managed ecosystems, and cities and settlements. AR5 WGIII assessed land-based climate change mitigation goals and pathways in the AFOLU. Here, this chapter assess land-climate interactions from all three working groups. It also builds on previous special reports such as the Special Report on Global Warming of 1.5°C (SR15). It links to the IPCC Guidelines on National Greenhouse Gas Inventories in the land sector. Importantly, this chapter assesses knowledge that has never been reported in any of those previous reports. Finally, the chapter also tries to reconcile the possible inconsistencies across the various IPCC reports.

Land-based water cycle changes: AR5 reported an increase in global evapotranspiration from the early 1980s to 2000s, but a constraint on further increases from low of soil moisture availability. Rising CO₂ concentration limits stomatal opening and thus also reduces transpiration, a component of evapotranspiration. Increasing aerosol levels, and declining surface wind speeds and levels of solar radiation reaching the ground are additional regional causes of the decrease in evapotranspiration.

Land area precipitation change: Averaged over the mid-latitude land areas of the Northern Hemisphere, precipitation has increased since 1901 (*medium confidence* before and *high confidence* after 1951). For other

1 latitudes, area-averaged long-term positive or negative trends have *low confidence*. There are *likely* more
2 land regions where the number of heavy precipitation events has increased than where it has decreased.
3 Extreme precipitation events over most of the mid-latitude land masses and over wet tropical regions will
4 very likely become more intense and more frequent (IPCC 2013a).

5
6 **Land-based GHGs:** AR5 reported that annual net CO₂ emissions from anthropogenic land use change were
7 0.9 [0.1–1.7] GtC yr⁻¹ on average during 2002 to 2011 (*medium confidence*). From 1750 to 2011, CO₂
8 emissions from fossil fuel combustion have released an estimated 375 [345–405] GtC to the atmosphere,
9 while deforestation and other land use change have released an estimated 180 [100–260] GtC. Of these
10 cumulative anthropogenic CO₂ emissions, 240 [230–250] GtC have accumulated in the atmosphere, 155
11 [125–185] GtC have been taken up by the ocean and 160 [70–250] GtC have accumulated in terrestrial
12 ecosystems (i.e., the cumulative residual land sink) (Ciais et al. 2013a). Updated assessment and knowledge
13 gaps are covered in Section 2.3.

14
15 **Future terrestrial carbon source/sink:** AR5 projected with *high confidence* that tropical ecosystems will
16 uptake less carbon and with *medium confidence* that at high latitudes, land carbon sink will increase in a
17 warmer climate. Thawing permafrost in the high latitudes is potentially a large carbon source at warmer
18 climate, but the magnitude of CO₂ and CH₄ emissions due to permafrost thawing is still uncertain. The SR15
19 further indicates that constraining warming to 1.5°C would prevent the melting of an estimated permafrost
20 area of 2 million km² over the next centuries compared to 2°C. Updates to these assessments are found in
21 Sections 2.3.

22
23 **Land use change altered albedo:** AR5 stated with *high confidence* that anthropogenic land use change has
24 increased the land surface albedo, which has led to a RF of $-0.15 \pm 0.10 \text{ W m}^{-2}$. However, it also underlined
25 that the sources of the large spread across independent estimates was caused by differences in assumptions
26 for the albedo of natural and managed surfaces and for the fraction of land use change before 1750.
27 Generally, our understanding of albedo changes from land use change has been enhanced from AR4 to AR5,
28 with a narrower range of estimates and a higher confidence level. The radiative forcing from changes in
29 albedo induced by land use changes was estimated in AR5 at -0.15 W m^{-2} (-0.25 to about -0.05), with
30 *medium confidence* in AR5 (Shindell et al. 2013). This was an improvement over AR4 in which it was
31 estimated at -0.2 W m^{-2} (-0.4 to about 0), with *low to medium confidence* (Forster et al. 2007). Section 2.5
32 shows that albedo is not the only source of biophysical land-based climate forcing to be considered.

33
34 **Hydrological feedback to climate:** Land use changes also affect surface temperatures through non-radiative
35 processes, and particularly through the hydrological cycle. These processes are less well known and are
36 difficult to quantify, but tend to offset the impact of albedo changes. As a consequence, there is low
37 agreement on the sign of the net change in global mean temperature as a result of land use change (Hartmann
38 et al. 2013a). An updated assessment on these points is covered in Section 2.5 and 2.2

39
40 **Climate-related extremes on land:** AR5 reported that impacts from recent climate-related extremes reveal
41 significant vulnerability and exposure of some ecosystems to current climate variability. Impacts of such
42 climate-related extremes include alteration of ecosystems, disruption of food production and water supply,
43 damage to infrastructure and settlements, morbidity and mortality, and consequences for mental health and
44 human well-being (Burkett et al. 2014). The SR15 further indicates that limiting global warming to 1.5°C
45 limits the risks of increases in heavy precipitation events in several regions (*high confidence*). In urban areas
46 climate change is projected to increase risks for people, assets, economies and ecosystems (*very high*
47 *confidence*). These risks are amplified for those lacking essential infrastructure and services or living in
48 exposed areas. Updated assessment and knowledge gap for this chapter are covered in Section 2.2 and Cross-
49 Chapter Box 4: Climate Change and Urbanisation.

50
51 **Land-based climate change adaptation and mitigation:** AR5 reported that adaptation and mitigation

1 choices in the near-term will affect the risks related to climate change throughout the 21st century (Burkett et
2 al. 2014). Agriculture, forestry and other land use (AFOLU) are responsible for about 10–12 GtCO₂eq yr⁻¹
3 anthropogenic greenhouse gas emissions mainly from deforestation and agricultural production. Global CO₂
4 emissions from forestry and other land use have declined since AR4, largely due to increased afforestation.
5 The SR15 further indicates that afforestation and bioenergy with carbon capture and storage (BECCS) are
6 important land-based carbon dioxide removal (CDR) options. It also states that land use and land-use change
7 emerge as a critical feature of virtually all mitigation pathways that seek to limit global warming to 1.5°C.
8 Climate Change 2014 Synthesis Report concluded that co-benefits and adverse side effects of mitigation
9 could affect achievement of other objectives such as those related to human health, food security,
10 biodiversity, local environmental quality, energy access, livelihoods and equitable sustainable development.
11 Updated assessment and knowledge gaps are covered in Section 2.6 and Chapter 7.

12
13 Overall, sustainable land management is largely constrained by climate change and extremes, but also puts
14 bounds on the capacity of land to effectively adapt to climate change and mitigate its impacts. Scientific
15 knowledge has advanced on how to optimise our adaptation and mitigation efforts while coordinating
16 sustainable land management across sectors and stakeholder. Details are assessed in subsequent sections.

17 **2.1.2 Introduction to the chapter structure**

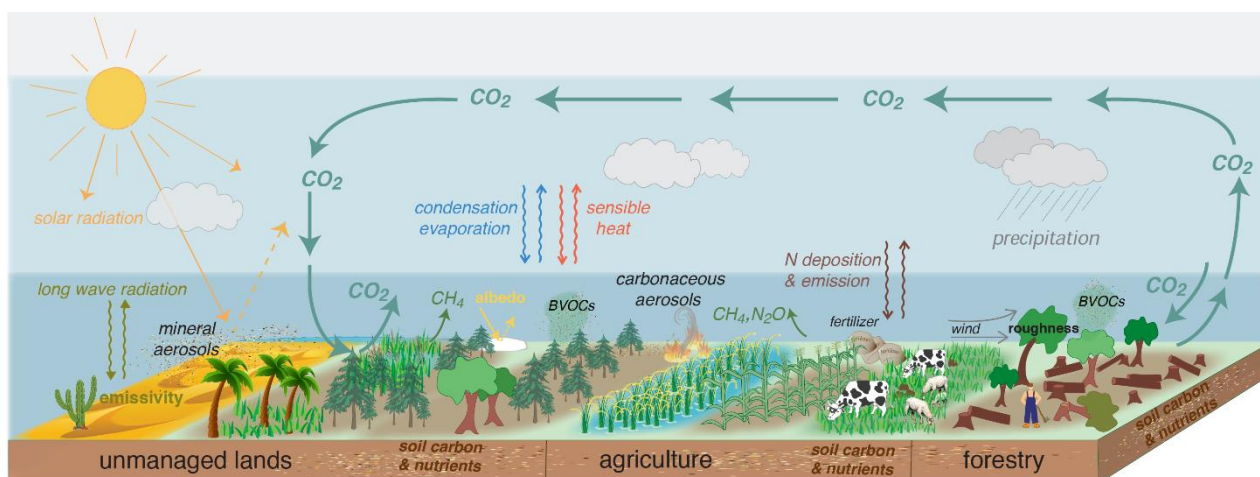
18
19
20 This chapter assess the consequences of changes in land cover and functioning, resulting from both land use
21 and climate change, to global and regional climates. The chapter starts by an assessment of the historical and
22 projected responses of land-based and processes to climate change and extremes (Section 2.2). Subsequently,
23 the chapter assesses historical and future changes in terrestrial GHG fluxes (Section 2.3), non-GHG fluxes
24 and precursors of SLCFs (Section 2.4. Section 2.4 focuses on how historical and future changes in land use
25 and land cover influence climate change/variability through biophysical and biogeochemical forcing and
26 feedbacks, how specific land management affects climate, and how in turn climate-induced land changes
27 feedback to climate. Section 2.6 assesses consequences of land-based adaptation and mitigation options for
28 the climate system in GHG and non-GHG exchanges. Sections 2.3 and 2.6 addresses implications of the
29 Paris Agreement for land-climate interactions, and the scientific evidence base for ongoing negotiations
30 around the Paris rulebook, the Global Stocktake, and credibility in measuring, reporting and verifying the
31 climate impacts of anthropogenic activities on land. The chapter also examines how land use and
32 management practices may affect climate change through biophysical feedbacks and radiative forcing
33 (Section 2.5), and assesses policy relevant projected land use changes and sustainable land management for
34 mitigation and adaptation (Section 2.6). Finally, the chapter concludes with a brief assessment of advances in
35 the understanding of ecological and biogeochemical processes underlying land-climate interactions (Section
36 2.7).

37
38 The chapter includes three chapter boxes providing general overview of (i) processes underlying land-
39 climate interactions (Box 2.1:); (ii) methodological approaches for estimating anthropogenic land carbon
40 fluxes from national to global scales (Box 2.2:); (iii) CO₂ fertilisation and enhanced terrestrial uptake of
41 carbon (Box 2.3). In addition this chapter includes two cross-chapter boxes on climate change and fire
42 (Cross-Chapter Box 3); and on urbanisation and climate change (Cross-Chapter Box 4).

43
44 In summary, the chapter assesses scientific understanding related to: 1) how a changing climate affects
45 terrestrial ecosystems, including those on managed lands; 2) how land affects climate through biophysical
46 and biogeochemical feedbacks; and 3) how land use or cover change and land management play an
47 important and complex role in the climate system. This chapter also pays special attention to advances in
48 understanding cross-scale interactions, emerging issues, heterogeneity, and teleconnections.

Box 2.1: Processes underlying land-climate interactions

Land continuously interacts with the atmosphere through exchanges of, for instance, greenhouse gases (e.g., CO₂, CH₄, N₂O), water, energy, or precursors of short lived-climate forcers (e.g., biogenic volatile organic compounds, dust, black carbon). The terrestrial biosphere also interacts with oceans through processes such as the influx of freshwater, nutrients, carbon and particles. These interactions affect where and when rain falls and thus irrigation needs for crops, frequency and intensity of heat waves, and air quality. They are modified by global and regional climate change, decadal, interannual and seasonal climatic variations, and weather extremes, as well as human actions on land (e.g., crop and forest management, afforestation and deforestation). This in turn affects atmospheric composition, surface temperature, hydrological cycle and thus local, regional and global climate. This box introduces some of the fundamental land processes governing biophysical and biogeochemical effects and feedbacks to the climate (Box 2.1 Figure 1)



Box 2.1 Figure 1 The structure and functioning of managed and unmanaged ecosystems that affect local, regional, and global climate. Land surface characteristics such as albedo and emissivity determine the amount of solar and long-wave radiation absorbed by land and reflected or emitted to the atmosphere. Surface roughness influences turbulent exchanges of momentum, energy, water, and biogeochemical tracers. Land ecosystems modulate the atmospheric composition through emissions and removals of many GHGs and precursors of SLCFs, including biogenic volatile organic compounds (BVOCs) and mineral dust. Atmospheric aerosols formed from these precursors affect regional climate by altering amounts of precipitation and of radiation reaching land surfaces through their role in clouds physics.

‘Biophysical interactions’ are exchanges of water and energy between the land and the atmosphere (Section 2.5). Land warms up from absorbing solar and long-wave radiation; it cools down through transfers of sensible heat (*via* conduction and convection) and latent heat (energy associated with water evapotranspiration) to the atmosphere and through longwave radiation emission from the land surface (Box 2.1 Figure 1). These interactions between the land and the atmosphere depend on the land surface characteristics, including reflectivity of short-wave radiation (albedo), emissivity of long wave radiation by vegetation and soils, surface roughness, and soil water access by vegetation, which depends on both soil characteristics and amounts of roots. Over seasonal, interannual and decadal time scales, these characteristics vary among different land cover and land-use types and are affected by both natural processes and land management (Anderson et al. 2011). A dense vegetation with high leaf area index, like forests, may absorb more energy than nearby herbaceous vegetation partly due to differences in surface albedo (especially when snow is on the ground). However, denser vegetation also sends more energy back to the atmosphere in the form of evapotranspiration (Bonan, 2008; Burakowski et al., 2018; Ellison et al., 2017; Section 2.5.2) and this contributes to changes in atmospheric water vapour content, affecting and rainfall.

Particularly in extra-tropical regions, these characteristics exhibit strong seasonal patterns with the development and senescence of the vegetation (e.g., leaf colour change and drop). For example, in deciduous forests, seasonal growth increases albedo by 20–50% from the spring minima to growing season maxima, followed by rapid decrease during leaf fall, whereas in grasslands, spring greening causes albedo decreases and only increases with vegetation browning (Hollinger et al. 2010). The seasonal patterns of sensible and

1 latent heat fluxes are also driven by the cycle of leaf development and senescence in temperate deciduous
2 forests: sensible heat fluxes peak in spring and autumn and latent heat fluxes peak in mid-summer (Moore et
3 al. 1996; Richardson et al. 2013).

4
5 Exchanges of greenhouse gases between the land and the atmosphere are referred to as ‘biogeochemical
6 interactions’ (Section 2.3), which are driven mainly by the balance between photosynthesis and respiration
7 by plants, and by the decomposition of soil organic matter by microbes. The conversion of atmospheric
8 carbon dioxide into organic compounds by plant photosynthesis, known as terrestrial net primary
9 productivity, is the source of plant growth, food for human and other organisms, and soil organic carbon.
10 Due to strong seasonal patterns of growth, northern hemisphere terrestrial ecosystems are largely responsible
11 for the seasonal variations in global atmospheric CO₂ concentrations. In addition to CO₂, soils emit
12 methane (CH₄) and nitrous oxide (N₂O) (Section 2.3). Soil temperature and moisture strongly affect
13 microbial activities and resulting fluxes of these three greenhouse gases.

14
15 Much like fossil fuel emissions, GHG emissions from anthropogenic land cover change and land
16 management are ‘forcers’ on the climate system. Other land-based changes to climate are described as
17 ‘feedbacks’ to the climate system - a process by which climate change influences some property of land,
18 which in turn diminishes (negative feedback) or amplifies (positive feedback) climate change. Examples of
19 feedbacks include the changes in the strength of land carbon sinks or sources, soil moisture and plant
20 phenology (Section 2.5.3).

21
22 Incorporating these land-climate processes into climate projections allows for increased understanding of the
23 land’s response to climate change (Section 2.2), and to better quantify the potential of land-based response
24 options for climate change mitigation (Section 2.6). However, to date Earth system models (ESMs)
25 incorporate some combined biophysical and biogeochemical processes only to limited extent and many
26 relevant processes about how plants and soils interactively respond to climate changes are still to be
27 included. (Section 2.7). And even within this class of models, the spread in ESM projections is large, in part
28 because of their varying ability to represent land-climate processes (Hoffman et al. 2014). Significant
29 progress in understanding of these processes has nevertheless been made since AR5.

2.2 The effect of climate variability and change on land

2.2.1 Overview of climate impacts on land

2.2.1.1 Climate drivers of land form and function

Energy is redistributed from the warm equator to the colder poles through large-scale atmospheric and oceanic processes driving the Earth's weather and climate (Oort and Peixoto 1983; Carissimo et al. 1985; Yang et al. 2015a). Subsequently, a number of global climate zones have been classified ranging from large-scale primary climate zones (tropical, sub-tropical, temperate, sub-polar, polar) to much higher-resolution, regional climate zones (e.g., the Köppen-Geiger classification, Kottek et al. 2006). Biomes are adapted to regional climates (Figure 2.1) and may shift as climate, land surface characteristics (e.g., geomorphology, hydrology), CO₂ fertilisation, and fire interact. These biomes and processes therein are subject to modes of natural variability in the ocean-atmosphere system that result in regionally wetter/drier or hotter/cooler periods having temporal scales from weeks to months (e.g., Southern Annular Mode), months to seasons (e.g., Madden-Julian Oscillation), years (e.g., El Niño Southern Oscillation) and decades (e.g. Pacific Decadal Oscillation). Furthermore, climate and weather extremes (such as drought, heat waves, very heavy rainfall, strong winds), whose frequency, intensity and duration are often a function of large-scale modes of variability, impact ecosystems at various space and time scales.

It is *very likely* that changes to natural climate variability as a result of global warming has and will continue to impact terrestrial ecosystems with subsequent impacts on land processes (Hulme et al. 1999; Parmesan and Yohe 2003; Di Lorenzo et al. 2008; Kløve et al. 2014; Berg et al. 2015; Lemordant et al. 2016; Pecl et al. 2017). This chapter assesses climate variability and change, particularly extreme weather and climate, in the context of desertification, land degradation, food security and terrestrial ecosystems more generally. This section does specifically assess the impacts of climate variability and climate change on desertification, land degradation and food security as these impacts are assessed respectively in Chapters 3, 4 and 5. This chapter begins with an assessment of observed warming on land.

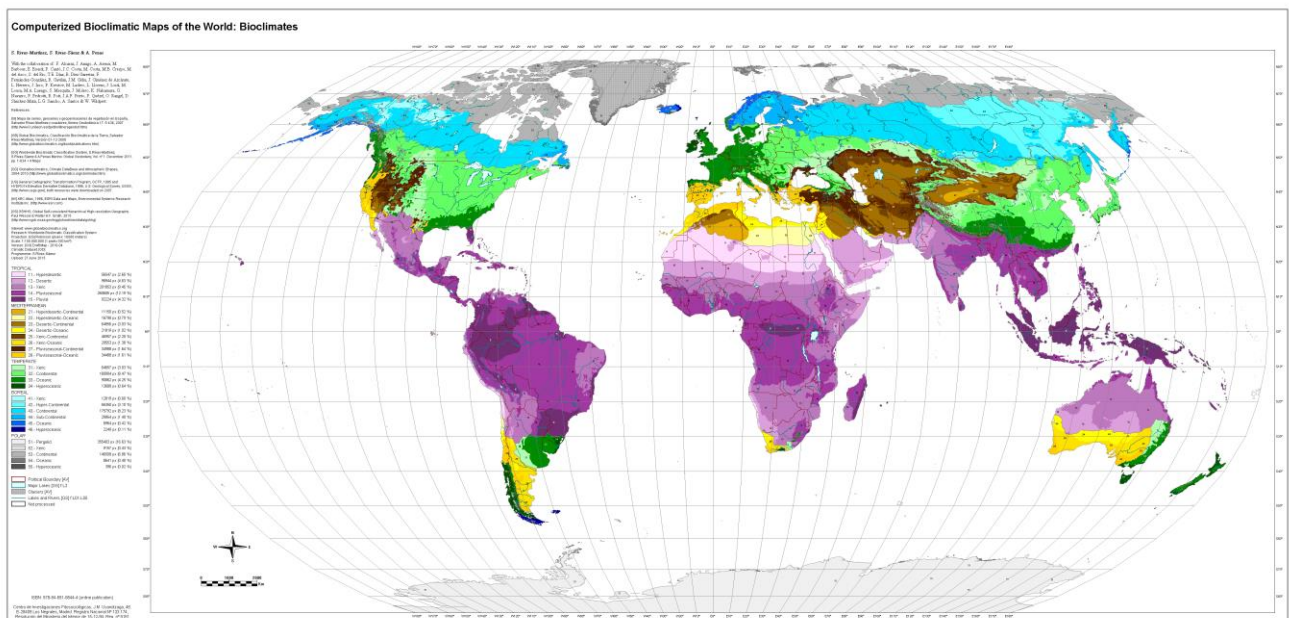


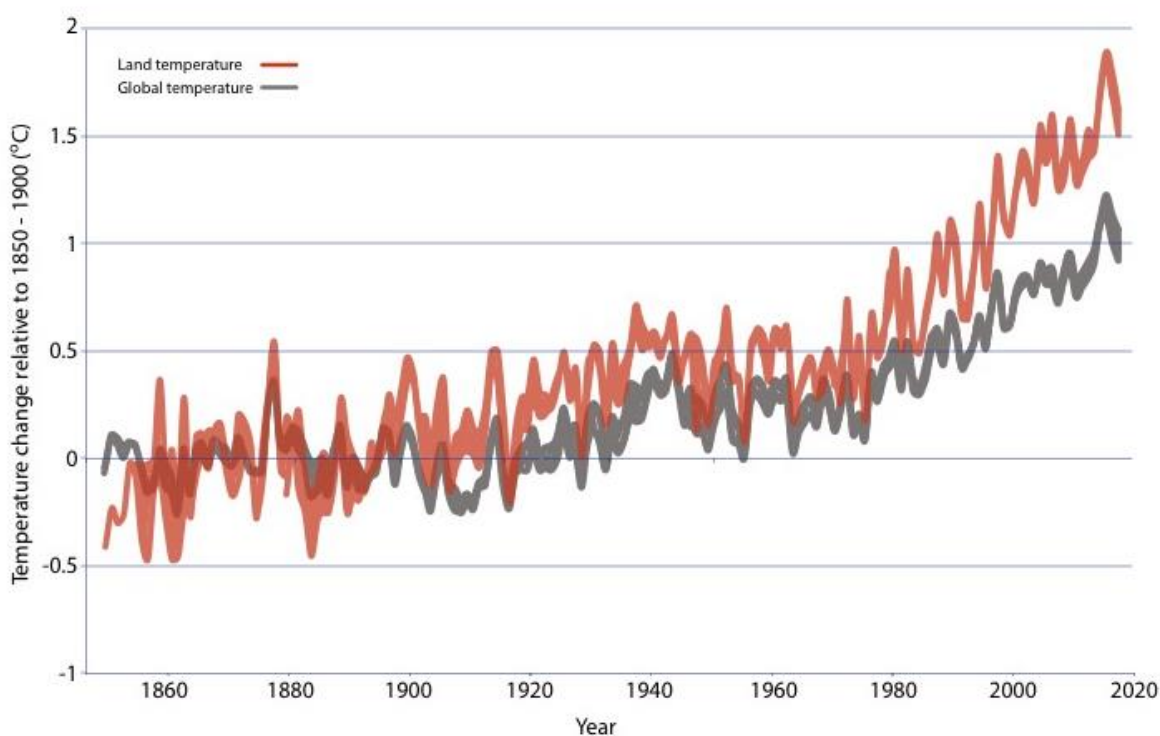
Figure 2.1 Worldwide Bioclimatic Classification System, 1996-2018. After (Rivas-Martinez et al. 2011).
Online at <http://www.globalbioclimatics.org>.

2.2.1.2 Changes in global land surface air temperature

Based on analysis of several global and regional land surface air temperature (LSAT) datasets, AR5 concluded that the global LSAT had increased over the instrumental period of record, with the warming rate approximately double that reported over the oceans since 1979 and that “it is certain that globally averaged LSAT has risen since the late 19th century and that this warming has been particularly marked since the

1 1970s". Warming found in the global land datasets is also in a broad agreement with station observations
 2 (Hartmann et al. 2013a).

3
 4 Since AR5, LSAT datasets have been improved and extended. The National Center for Environmental
 5 Information, which is a part of the US National Oceanic and Atmospheric Administration (NOAA),
 6 developed a new version of the Global Historical Climatology Network (GHCNm, version 4) dataset. The
 7 dataset provides an expanded set of station temperature records with more than 25,000 total monthly
 8 temperature stations compared to 7200 in versions v2 and v3 (Menne et al. 2018). Goddard Institute for
 9 Space Studies, which is a part of the US National Aeronautics and Space Administration, (NASA/GISS)
 10 provides estimate of land and ocean temperature anomalies (GISTEMP). The GISTEMP land temperature
 11 anomalies are based upon primarily NOAA/GHCN version 3 dataset (Lawrimore et al. 2011) and account for
 12 urban effects through nighttime adjustments (Hansen et al. 2010). The Climatic Research Unit of the
 13 University of East Anglia, UK (CRUTEM) dataset, now version CRUTEM4.6, incorporates additional
 14 stations (Jones et al. 2012). Finally, the Berkeley Earth Surface Temperature (BEST) dataset provides LSAT
 15 from 1750 to present based on almost 46,000 time series and has the longest temporal coverage of the four
 16 datasets (Rohde et al. 2013). This dataset was derived with methods distinct from those used for
 17 development of the NOAA and NASA datasets and the CRU dataset.
 18
 19



20
 21
 22 **Figure 2.2 Evolution of land surface air temperature (LSAT) and global mean surface temperature**
 23 **(GMST) over the period of instrumental observations. Red line shows annual mean LSAT in the**
 24 **Berkeley, CRUTEM4, GHCNv4 and GISTEMP datasets, expressed as departures from global average**
 25 **LSAT in 1850–1900, with the red line thickness indicating inter-dataset range. Gray shaded line shows**
 26 **annual mean Global Mean Surface Temperature (GMST) in the HadCRUT4, NOAA Global Temp,**
 27 **GISTEMP and Cowtan&Way datasets (monthly values of which were reported in the Special Report on**
 28 **Global Warming of 1.5°C (Allen et al. 2018)).**
 29

30 According to the available observations in the four datasets, the globally averaged LSAT increased by
 31 1.44°C from the preindustrial period (1850–1900) to present (1999–2018). The warming from the late 19th
 32 century (1881–1900) to present (1999–2018) was 1.41°C (1.31°C–1.51°C) (Table 2.1). The 1.31°C–1.51°C
 33 range represents the spread in median estimates from the four available land datasets and does not reflect
 34 uncertainty in data coverage or methods used. Based on the Berkeley dataset (the longest dataset with the
 35 most extensive land coverage) the total increase in LSAT between the average of the 1850–1900 period and

1 the 1999–2018 period was 1.52°C, (1.39°C–1.66°C; 95% confidence).

2 The extended and improved land datasets reaffirmed the AR5 conclusion that it is certain that globally
3 averaged LSAT has risen since the preindustrial period and that this warming has been particularly marked
4 since the 1970s (Figure 2.2).

5
6 **Table 2.1 Increases in land surface air temperature (LSAT) from preindustrial
7 period and the late 19th century to present (1999–2018).**

Reference period	Dataset of LSAT increase (C°)			
	Berkeley	CRUTEM4	GHCNm, v4	GISTEMP
<i>Preindustrial</i> 1850–1900	1.52 1.39–1.66 (95% confidence)	1.31	NA	NA
<i>Late 19th century</i> 1881–1900	1.51 1.40–1.63 (95% confidence)	1.31	1.37	1.45

8
9 Recent analyses of LSAT and sea surface temperature (SST) observations as well as analyses of climate
10 model simulations have refined our understanding of underlying mechanisms responsible for a faster rate of
11 warming over land than over oceans. Analyses of paleo records, historical observations, model simulations,
12 and underlying physical principles are all in agreement that that land is warming faster than the oceans as a
13 result of differences in evaporation, land-climate feedbacks (Section 2.5), and changes in the aerosol forcing
14 over land (Braconnot et al. 2012; Joshi et al. 2013; Sejas et al. 2014; Byrne and O’Gorman 2013, 2015;
15 Wallace and Joshi 2018; Allen et al. 2019) (*very high confidence*). There is also *high confidence* that
16 difference in land and ocean heat capacity is not the primary reason for a faster land than ocean warming.
17 For the recent period the land-to-ocean warming ratio is in close agreement between different observational
18 records (about 1.6) and the CMIP5 climate model simulations (the *likely* range of 1.54°C to 1.81°C). Earlier
19 studies analysing slab ocean models (models in which it is assumed that the deep ocean has equilibrated)
20 produced a higher land temperature increases than sea surface temperature (Manabe et al. 1991; Sutton et al.
21 2007).

22
23 It is certain that globally averaged LSAT has risen faster than GMST from preindustrial (1850–1900) to
24 present day (1999–2018). This is because the warming rate of the land compared to the ocean is substantially
25 higher over the historical period (by approximately 60%) and because Earth surface is approximately one
26 third land and two thirds ocean. This enhanced land warming impacts land processes with implications for
27 desertification (Section 2.2.2, Chapter 3), food security (Section 2.2.3, Chapter 5), terrestrial ecosystems
28 (Section 2.2.4), and GHG and non-GHG fluxes between the land and climate (Section 2.3, Section 2.4).
29 Future changes in land characteristics through adaptation and mitigation processes and associated land-
30 climate feedbacks can dampen warming in some regions and enhance warming in others (Section 2.5).

31 32 **2.2.2 Climate driven changes in aridity**

33
34 Desertification is defined and discussed at length in Chapter 3 of this report and is a function of both human
35 activity and climate variability and change. There are uncertainties in distinguishing between historical
36 climate-caused aridification and desertification and also future projections of aridity as different
37 measurement methods of aridity do not agree on historical or projected changes (3.1.1, 3.2.1). However,
38 warming trends over drylands are twice the global average (Lickley and Solomon 2018) some temperate
39 drylands are projected to convert to subtropical drylands as a result of an increased drought frequency
40 causing reduced soil moisture availability in the growing season (Engelbrecht et al. 2015; Schlaepfer et al.
41 2017). We therefore assess with *medium confidence* that a warming climate will result in regional increases
42 in the spatial extent of drylands under mid- and high emission scenarios and that these regions will warm
43 faster than the global average warming rate.

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2.2.3 The influence of climate change on food security

Food security and the various components thereof is addressed in depth in Chapter 5. Climate variables relevant to food security and food systems are predominantly temperature and precipitation-related, but also include integrated metrics that combine these and other variables (like solar radiation, wind, humidity) and extreme weather and climate events including storm surge (see 5.2.1). The impact of climate change through changes in these variables is projected to negatively impact all aspects of food security (food availability, access, utilisation and stability), leading to complex impacts on global food security (Chapter 5) (Table 5.1) (*high confidence*).

Climate change will have regionally distributed impacts, even under aggressive mitigation scenarios (Howden et al. 2007; Rosenzweig et al. 2013; Challinor et al. 2014; Parry et al. 2005; Lobell and Tebaldi 2014; Wheeler and Von Braun 2013). For example, in the northern hemisphere the northward expansion of warmer temperatures in the middle and higher latitudes will lengthen the growing season (Gregory and Marshall 2012; Yang et al. 2015b) which may benefit crop productivity (Parry et al. 2004; Rosenzweig et al., 2014; Deryng et al. 2016). However, continued rising temperatures are expected to impact global wheat yields by about 4–6% reductions for every degree of temperature rise (Liu et al. 2016a; Asseng et al. 2015) and across both mid- and low latitude regions, rising temperatures are also expected to be a constraining factor for maize productivity by the end of the century (Bassu et al. 2014; Zhao et al. 2017). Although there has been a general reduction in frost occurrence during winter and spring and a lengthening of the frost free season in response to growing concentrations of greenhouse gases (Fischer and Knutti 2014; Wypych et al. 2017), there are regions where the frost season length has increased e.g. southern Australia (Crimp et al. 2016). Despite the general reduced frost season length, late spring frosts may increase risk of damage to warming induced precocious vegetation growth and flowering (Meier et al. 2018). Observed and projected warmer minimum temperatures have and will continue to reduce the number of winter chill units required by particularly fruit crops (Luedeling 2012). Crop yields are impacted negatively by increases of seasonal rainfall variability in the tropics, sub-tropics, water-limited and high elevation environments, drought severity and growing season temperatures have a negative impact on crop yield (IFPRI 2009; Schlenker and Lobell 2010; Müller et al. 2017; Parry et al. 2004; Wheeler and Von Braun 2013; Challinor et al. 2014).

Changes in extreme weather and climate (Section 2.2.5) have negative impacts on food security through regional reductions of crop yields. A recent study shows that between 18-43% of the explained yield variance of four crops (maize, soybeans, rice and spring wheat) is attributable to extremes of temperature and rainfall, depending on the crop type (Vogel et al. 2019). Climate shocks, particularly severe drought impact low-income small-holder producers disproportionately (Vermeulen et al. 2012b; Rivera Ferre 2014). Extremes also compromise critical food supply chain infrastructure, making the transport and access to harvested food more difficult (Brown et al. 2015; Fanzo et al. 2018). There is *high confidence* that the impacts of enhanced climate extremes, together with non-climate factors such as nutrient limitation, soil health and competitive plant species, generally outweighs the regionally positive impacts of warming (Lobell et al. 2011; Leakey et al. 2012; Porter et al. 2014; Gray et al. 2016; Pugh et al. 2016; Wheeler and Von Braun 2013; Beer 2018).

2.2.4 Climate-driven changes in terrestrial ecosystems

Previously, the IPCC AR5 reported high confidence that the Earth's biota composition and ecosystem processes have been strongly affected by past changes in global climate, but the rates of the historic climate change are lower than those projected for the 21st century under high warming scenarios like RCP8.5

1 (Settele et al. 2015a). There is *high confidence* that as a result of climate changes over recent decades many
2 plant and animal species have experienced range size and location changes, altered abundances, and shifts in
3 seasonal activities (Urban 2015a; Ernakovich et al. 2014; Elsen and Tingley 2015; Hatfield and Prueger
4 2015; Urban 2015b; Savage and Vellend 2015; Yin et al. 2016; Pecl et al. 2017; Gonsamo et al. 2017;
5 Fadrique et al. 2018; Laurance et al. 2018). There is high confidence that climate zones have already shifted
6 in many parts of the world primarily as an increase of dry, arid climates accompanied by a decrease of polar
7 climates (Chan and Wu 2015; Chen and Chen 2013; Spinoni et al. 2015b). Regional climate zones shifts
8 have been observed over the Asian monsoon region (Son and Bae 2015), Europe (Jylhä et al. 2010), China
9 (Yin et al. 2019), Pakistan (Adnan et al. 2017), the Alps (Rubel et al. 2017) and North-Eastern Brazil,
10 Southern Argentina, the Sahel, Zambia and Zimbabwe, the Mediterranean area, Alaska, Canada and North-
11 Eastern Russia (Spinoni et al. 2015b).

12
13 There is *high confidence* that bioclimates zones will further shift as the climate warms (Williams et al. 2007;
14 Rubel and Kottek 2010; Garcia et al. 2016; Mahony et al. 2017; Law et al. 2018). There is also high
15 confidence that novel, unprecedented climates (climate conditions with no analog in the observational
16 record) will emerge, particularly the tropics (Williams and Jackson 2007; Colwell et al. 2008a; Mora et al.
17 2013, 2014; Hawkins et al. 2014; Mahony et al. 2017; Maule et al. 2017). It is *very likely* that terrestrial
18 ecosystems and land processes will be exposed to disturbances beyond the range of current natural
19 variability as a result of global warming, even under low- to medium-range warming scenarios, and these
20 disturbances will alter the structure, composition and functioning of the system (Settele et al. 2015b;
21 Gauthier et al. 2015; Seddon et al. 2016).

22
23 In a warming climate many species will be unable to track their climate niche as it moves, especially those in
24 extensive flat landscapes with low dispersal capacity and in the tropics whose thermal optimum is already
25 near current temperature (Diffenbaugh and Field 2013; Warszawski et al. 2013). Range expansion in higher
26 latitudes and elevations as a result of warming often, but not exclusively occurs in abandoned lands (Harsch
27 et al. 2009; Landh usser et al. 2010; Gottfried et al. 2012; Boisvert-Marsh et al. 2014; Bryn and Potthoff
28 2018; Rumpf et al. 2018; Buitenwerf et al. 2018; Steinbauer et al. 2018). This expansion typically favours
29 thermophilic species at the expense of cold adapted species as climate becomes suitable for lower
30 latitude/altitude species (Rumpf et al. 2018). In temperate drylands, however, range expansion can be
31 countered by intense and frequent drought conditions which result in accelerated rates of taxonomic change
32 and spatial heterogeneity in an ecotone (Tietjen et al. 2017).

33
34 Since the advent of satellite observation platforms, a global increase in vegetation photosynthetic activity
35 (i.e. greening) as evidenced through remotely sensed indices such as leaf area index (LAI) and normalised
36 difference vegetation index (NDVI). Three satellite-based leaf area index (GIMMS3g, GLASS and
37 GLOMAP) records imply increased growing season LAI (greening) over 25–50% and browning over less
38 than 4% of the global vegetated area, resulting in greening trend of $0.068 \pm 0.045 \text{ m}^2 \text{ m}^{-2} \text{ yr}^{-1}$ over 1982–2009
39 (Cao et al. 2016). Greening has been observed in southern Amazonia, southern Australia, the Sahel and
40 central Africa, India, eastern China and the northern extratropical latitudes (Myneni et al. 1997; de Jong et al.
41 2012; Los 2013; Piao et al. 2015; Mao et al. 2016; Zhu et al. 2016; Carlson et al. 2017; Forzieri et al. 2017;
42 Pan et al. 2018; Chen et al. 2019). Greening has been attributed to direct factors, namely human land use
43 management and indirect factors such as CO₂ fertilisation, climate change, nitrogen deposition (Donohue et
44 al. 2013; Keenan et al. 2016; Zhu et al. 2016). Indirect factors have been used to explain most greening
45 trends primarily through CO₂ fertilisation in the tropics and through an extended growing season and
46 increased growing season temperatures as a result of climate change in the high latitudes (Fensholt et al.
47 2012; Zhu et al. 2016). The extension of the growing season in high latitudes has occurred together with an
48 earlier spring greenup (the time at which plants begin to produce leaves in northern mid- and high-latitude
49 ecosystems) (Goetz et al. 2015; Xu et al. 2016a, 2018) with subsequent earlier spring carbon uptake (2.3
50 days per decade) and gross primary productivity (GPP) (Pulliainen et al. 2017). The role of direct factors of

1 greening are being increasingly investigated and a recent study has attributed over a third of observed global
2 greening between 2000 and 2017 to direct factors, namely afforestation and croplands, in China and India
3 (Chen et al. 2019).

4
5 It should be noted that as measured greening is a product of satellite-derived radiance data, and as such does
6 not provide information on ecosystem health indicators such as species composition and richness,
7 homeostasis, absence of disease, vigor, system resilience and the different components of ecosystems
8 (Jørgensen et al. 2016). For example, a regional greening attributable to croplands expansion or
9 intensification might occur at the expense of ecosystem biodiversity.

10
11 Within the global greening trend are also detected regional decreases in vegetation photosynthetic activity
12 (i.e. browning) in northern Eurasia, the southwestern USA, boreal forests in North America, Inner Asia and
13 the Congo Basin, largely as a result of intensified drought stress. Since the late-1990s rates and extents of
14 browning have exceeded those of greening in some regions, the collective result of which has been a
15 slowdown of the global greening rate (de Jong et al. 2012; Pan et al. 2018). Within these long-term trends,
16 interannual variability of regional greening and browning is attributable to regional climate variability,
17 responses to extremes such as drought, disease and insect infestation and large-scale teleconnective controls
18 such as ENSO and the Atlantic Multi-decadal Organization (Verbyla 2008; Revadekar et al. 2012; Epstein et
19 al. 2018; Zhao et al. 2018).

20
21 Projected increases in drought conditions in many regions suggest long-term global vegetation greening
22 trends are at risk of reversal to browning in a warmer climate (de Jong et al. 2012; Pan et al. 2018; Pausas
23 and Millán 2018). On the other hand, in higher latitudes vegetation productivity is projected to increase as a
24 result of higher atmospheric CO₂ concentrations and longer growing periods as a result of warming here (Ito
25 et al. 2016)(Section 2.3, Box 2.3). Additionally, climate-driven transitions of ecosystems, particularly range
26 changes, can take years to decades for the equilibrium state to be realised and the rates of these “committed
27 ecosystem changes” (Jones et al. 2009) vary between low and high latitudes (Jones et al. 2010). Furthermore,
28 as direct factors are poorly integrated into Earth systems models (ESMs) uncertainties in projected trends of
29 greening and browning are further compounded (Buitenwerf et al. 2018; Chen et al. 2019). Therefore, there
30 is *low confidence* in the projection of global greening and browning trends.

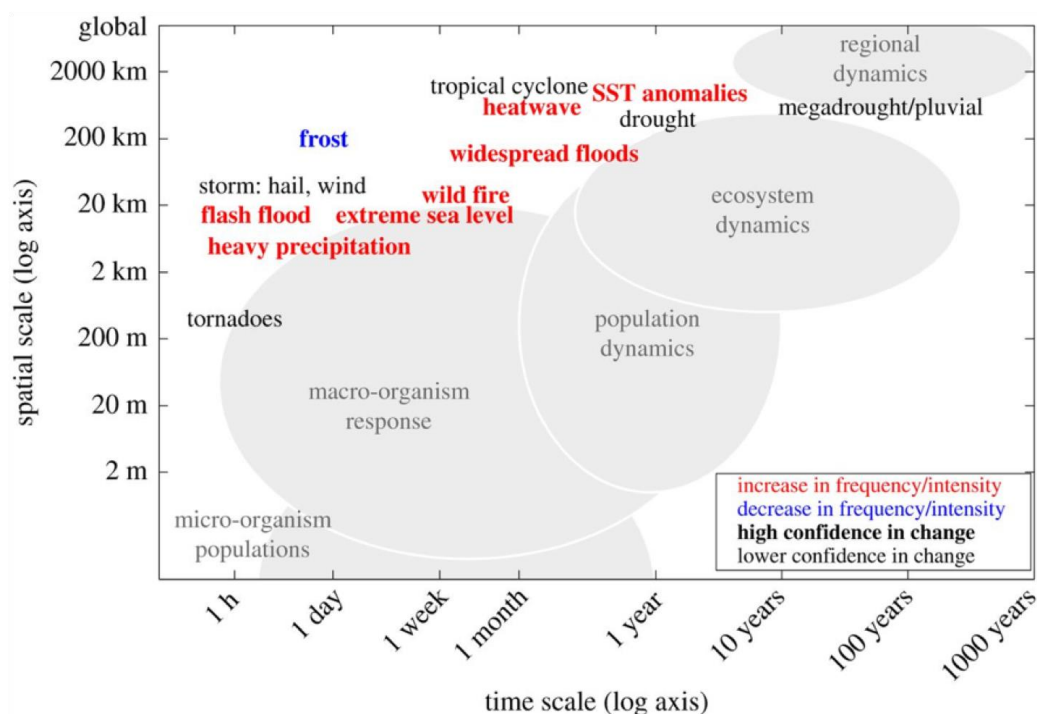
31
32 Increased atmospheric CO₂ concentrations have both direct and indirect effects on terrestrial ecosystems (see
33 Sections 2.2.2 and 2.2.3, Box 2.3). The direct effect is primarily through increased vegetation photosynthetic
34 activity as described above. Indirect effects include decreased evapotranspiration that may offset the
35 projected impact of drought in some water-stressed plants through improved water use efficiency in
36 temperate regions suggesting that some rain-fed cropping systems and grasslands will benefit from elevated
37 atmospheric CO₂ concentrations (Roy et al. 2016a; Milly and Dunne 2016; Swann et al. 2016; Chang et al.
38 2017; Zhu et al. 2017). In tropical regions increased flowering activity is associated primarily with
39 increasing atmospheric CO₂ suggesting a long-term increase in flowering activity may persist in some
40 vegetation, particularly mid-story trees and tropical shrubs, and enhance reproduction levels until limited by
41 nutrient availability or climate factors like drought frequency, rising temperatures and reduced insolation
42 (Pau et al. 2018).

43 44 **2.2.5 Climate extremes and their impact on land functioning**

45
46 Extreme weather events are generally defined as the upper or lower statistical tails of the observed range of
47 values of climate variables or climate indicators (e.g., temperature/rainfall or drought/aridity indices
48 respectively). Previous IPCC reports have reported with *high confidence* on the increase of many types of
49 observed extreme temperature events (Seneviratne et al. 2012b; Hartmann et al. 2013b; Hoegh-Guldberg et
50 al. 2018). However, as a result of observational constraints, increases in precipitation extremes are less

1 confident, except in observation rich regions with dense, long-lived station networks such as Europe and
 2 North America where there have been likely increases in the frequency or intensity of heavy rainfall.

3
 4 Extreme events occur across a wide range of time and space scales (Figure 2.3) and may include individual,
 5 relatively short-lived weather events (e.g., extreme thunderstorms storms) or a combination or accumulation
 6 of non-extreme events (Colwell et al. 2008b; Kundzewicz and Germany 2012) e.g., moderate rainfall in a
 7 saturated catchment having the flood peak at mean high tide (Leonard et al. 2014). Combinatory processes
 8 leading to a significant impact are referred to as a compound event and are a function of the nature and
 9 number of physical climate and land variables, biological agents such as pests and disease, the range of
 10 spatial and temporal scales, the strength of dependence between processes, and the perspective of the
 11 stakeholder who defines the impact (Leonard et al. 2014; Millar and Stephenson 2015). Current *confidence*
 12 in the impact of compound events on land is *low* as the multi-disciplinary approaches needed to address the
 13 problem are few (Zscheischler et al. 2018) and the rarity of compound extreme climatic events renders the
 14 analysis of impacts difficult.



16
 17
 18 **Figure 2.3** Spatial and temporal scales of typical extreme weather and climate events and the biological
 19 systems they impact (shaded grey). Individuals, populations and ecosystems within these space-time
 20 ranges respond to relevant climate stressors. Red (blue) labels indicate an increase (decrease) in the
 21 frequency or intensity of the event, with bold font reflecting confidence in the change. Non-bold black
 22 labels indicate low confidence in observed changes in frequency or intensity of these events. Each event
 23 type indicated in the figure is likely to affect biological systems at all temporal and spatial scales located
 24 to the left and below the specific event position in the figure. From Ummerhofer and Meehl (2017).

25 26 2.2.5.1 Changes in extreme temperatures, heat waves and drought

27 It is *very likely* that most land areas have experienced a decrease in the number of cold days and nights, and
 28 an increase in the number of warm days and unusually hot nights (Orlowsky and Seneviratne 2012;
 29 Seneviratne et al. 2012; Mishra et al. 2015; Ye et al. 2018). Although there is no consensus definition of heat
 30 waves as some heat wave indices have relative thresholds and others absolute thresholds, trends between
 31 indices of the same type show that recent heat-related events have been made more frequent or more intense
 32 due to anthropogenic greenhouse gas emissions in most land regions (Lewis and Karoly 2013; Smith et al.
 33 2013b; Scherer and Diffenbaugh 2014; Fischer and Knutti 2015; Ceccherini et al. 2016; King et al. 2016;

1 Bador et al. 2016; Stott et al. 2016; King 2017; Hoegh-Guldberg et al. 2018). Globally, 50–80 % of the land
2 fraction is projected to experience significantly more intense hot extremes than historically recorded (Fischer
3 and Knutti 2014; Diffenbaugh et al. 2015; Seneviratne et al. 2016). There is *high confidence* that heat waves
4 will increase in frequency, intensity and duration into the 21st century (Russo et al. 2016; Ceccherini et al.
5 2017; Herrera-Estrada and Sheffield 2017) and under high emission scenarios heat waves by the end of the
6 century may become extremely long (more than 60 consecutive days) and frequent (once every two years) in
7 Europe, North America, South America, Africa, Indonesia, the Middle East, south and south east Asia and
8 Australia (Rusticucci 2012; Cowan et al. 2014; Russo et al. 2014; Scherer and Diffenbaugh 2014; Pal and
9 Eltahir 2016; Rusticucci et al. 2016; Schär 2016; Teng et al. 2016; Dosio 2017; Mora et al. 2017; Dosio et al.
10 2018; Lehner et al. 2018; Lhotka et al. 2018; Lopez et al. 2018; Tabari and Willems 2018). Furthermore,
11 unusual heat wave conditions today will occur regularly by 2040 under the RCP 8.5 scenario (Russo et al.
12 2016). The intensity of heat events may be modulated by the land cover and soil characteristics (Miralles et
13 al. 2014; Lemordant et al. 2016; Ramarao et al. 2016). Where temperature increase results in decreased soil
14 moisture, latent heat flux is reduced while sensible heat fluxes is increased allowing surface air temperature
15 to rise further. However, this feedback may be diminished if the land surface is irrigated through the
16 enhanced evapotranspiration (Mueller et al. 2015; Siebert et al. 2017)(Section 2.5.2.2).

17
18 Drought (Qin et al. 2013), including megadroughts of the last century, e.g., the Dustbowl drought (Hegerl et
19 al. 2018), Chapter 5), is a normal component of climate variability (Hoerling et al. 2010; Dai 2011) and may
20 be seasonal, multi-year (Van Dijk et al. 2013) or multi-decadal (Hulme 2001) with increasing degrees of
21 impact on the regional activities. This interannual variability is controlled particularly through remote sea
22 surface temperature (SST) forcings such as the Inter-decadal Pacific Oscillation (IPO) and the Atlantic
23 Multi-decadal Oscillation (AMO), El Niño/Southern Oscillation (ENSO) and Indian Ocean Dipole (IOD),
24 that cause drought as a result of reduced rainfall (Kelley et al. 2015; Dai 2011; Hoell et al. 2017; Espinoza et
25 al. 2018). In some cases however, large scale SST modes do not fully explain the severity of drought some
26 recent event attribution studies have identified a climate change fingerprint in several regional droughts, e.g.,
27 the western Amazon (Erfanian et al. 2017), southern Africa (Funk et al. 2018; Yuan et al. 2018), southern
28 Europe and the Mediterranean including North Africa (Kelley et al. 2015; Wilcox et al. 2018), parts of North
29 America (Williams et al. 2015; Mote et al. 2016), Russia (Otto et al. 2012), India (Ramarao et al. 2015) and
30 Australia (Lewis and Karoly 2013).

31
32 Long-term global trends in drought are difficult to determine because of this natural variability, potential
33 deficiencies in drought indices (especially in how evapotranspiration is treated) and the quality and
34 availability of precipitation data (Sheffield et al. 2012; Dai 2013; Trenberth et al. 2014; Nicholls and
35 Seneviratne 2015; Mukherjee et al. 2018). However, regional trends in frequency and intensity of drought
36 are evident in several parts of the world, particularly in low latitude land areas, such as the Mediterranean,
37 North Africa and Middle East (Vicente-Serrano et al. 2014; Spinoni et al. 2015a; Dai and Zhao 2017; Páscoa
38 et al. 2017), many regions of sub-Saharan Africa (Masih et al. 2014; Dai and Zhao 2017), Central China
39 (Wang et al. 2017e), the southern Amazon (Fu et al. 2013; Espinoza et al. 2018), India (Ramarao et al.
40 2016), east and south Asia, parts of North America and eastern Australia (Dai and Zhao 2017). A recent
41 analysis of 4500 meteorological droughts globally found increased drought frequency over the U.S. East
42 Coast, Amazonia and north-eastern Brazil, Patagonia, the Mediterranean region, most of Africa and north-
43 eastern China with decreased drought frequency over northern Argentina, Uruguay and northern Europe
44 (Spinoni et al. 2019). The study also found drought intensity has become more severe over north-western
45 U.S., parts of Patagonia and southern Chile, the Sahel, the Congo River basin, southern Europe, north-
46 eastern China, and south-eastern Australia, whereas the eastern U.S., south-eastern Brazil, northern Europe,
47 and central-northern Australia experienced less severe droughts. In addition to the IPCC SR15 assessment of
48 medium confidence in increased drying over the Mediterranean region (Hoegh-Guldberg et al. 2018), It is
49 further assessed with *medium confidence* an increased frequency and intensity of drought in Amazonia and
50 north-eastern Brazil, Patagonia, most of Africa and north-eastern China.

1 There is *low confidence* on how large-scale modes of variability will respond to a warming climate (Deser et
2 al. 2012; Liu 2012; Christensen et al. 2013; Hegerl et al. 2015; Newman et al. 2016). Although, there is
3 evidence for an increased frequency of extreme ENSO events, such as the 1997/98 El Niño and 1988/89 La
4 Niña (Cai et al. 2014a, 2015) and extreme positive phases of the IOD (Christensen et al. 2013; Cai et al.
5 2014b). However, The assessment by the SR15 was retained on an increased regional drought risk (*medium*
6 *confidence*), specifically over the Mediterranean and South Africa at both 1.5°C and 2°C warming levels
7 compared to present day, with drought risk at 2°C being significantly higher than at 1.5°C (Hoegh-Guldberg
8 et al. 2018).

10 **2.2.5.2 Impacts of heat extremes and drought on land**

11 There is *high confidence* that heat extremes such as unusually hot nights, extremely high daytime
12 temperatures, heat waves and drought are damaging to crop production (Chapter 5). Extreme heat events
13 impact a wide variety of tree functions including reduced photosynthesis, increased photooxidative stress,
14 leaves abscise, a decreased growth rate of remaining leaves and decreased growth of the whole tree (Teskey
15 et al. 2015). Although trees are more resilient to heat stress than grasslands (Teuling et al. 2010), it has been
16 observed that different types of forest (e.g., needleleaf vs. broadleaf) respond differently to drought and heat
17 waves (Babst et al. 2012). For example, in the Turkish Anatolian forests net primary productivity (NPP)
18 generally decreased during drought and heat waves events between 2000 and 2010 but in a few other
19 regions, NPP of needle leaf forests increased (Erşahin et al. 2016). However, forests may become less
20 resilient to heat stress in future due to the long recovery period required to replace lost biomass and the
21 projected increased frequency of heat and drought events (Frank et al. 2015a; McDowell and Allen 2015;
22 Johnstone et al. 2016; Stevens-Rumann et al. 2018). Additionally, widespread regional tree mortality may be
23 triggered directly by drought and heat stress (including warm winters) and exacerbated by insect outbreak
24 and fire (Neuvonen et al. 1999; Breshears et al. 2005; Berg et al. 2006; Soja et al. 2007; Kurz et al. 2008b;
25 Allen et al. 2010) .

26
27 Gross primary production (GPP) and soil respiration form the first and second largest carbon fluxes from
28 terrestrial ecosystems to the atmosphere in the global carbon cycle (Beer et al. 2010; Bond-Lamberty and
29 Thomson 2010). Heat extremes impact the carbon cycle through altering these and change ecosystem-
30 atmosphere CO₂ fluxes and the ecosystem carbon balance. Compound heat and drought events result in a
31 stronger carbon sink reduction compared to single-factor extremes as GPP is strongly reduced and ecosystem
32 respiration less so (Reichstein et al. 2013; Von Buttlar et al. 2018). In forest biomes, however, GPP may
33 increase temporarily as a result of increased insolation and photosynthetic activity as was seen during the
34 2015-2016 ENSO related drought over Amazonia (Zhu et al. 2018). Longer extreme events (heat wave or
35 drought or both) result in a greater reduction in carbon sequestration and may also reverse long-term carbon
36 sinks (Ciais et al. 2005; Phillips et al. 2009; Wolf et al. 2016b; Ummenhofer and Meehl 2017; Von Buttlar et
37 al. 2018; Reichstein et al. 2013). Furthermore, extreme heat events may impact the carbon cycle beyond the
38 lifetime of the event. These lagged effects can slow down or accelerate the carbon cycle: it will slow down if
39 reduced vegetation productivity and/or widespread mortality after an extreme drought are not compensated
40 by regeneration, or speed up if productive tree and shrub seedlings cause rapid regrowth after windthrow or
41 fire (Frank et al. 2015a). Although some ecosystems may demonstrate resilience to a single heat climate
42 stressor like drought, e.g. forests, compound effects of, e.g., deforestation, fire and drought potentially can
43 result in changes to regional precipitation patterns and river discharge, losses of carbon storage and a
44 transition to a disturbance-dominated regime (Davidson et al. 2012). Additionally, adaptation to seasonal
45 drought may be overwhelmed by multi-year drought and their legacy effects (Brando et al. 2008; da Costa et
46 al. 2010).

47
48 Under medium and high emission scenarios, global warming will exacerbate heat stress thereby amplifying
49 deficits in soil moisture and runoff despite uncertain precipitation changes (Ficklin and Novick 2017; Berg
50 and Sheffield 2018; Cook et al. 2018; Dai et al. 2018; Engelbrecht et al. 2015; Ramarao et al. 2015; Grillakis

2019). This will increase the rate of drying causing drought to set in quicker, become more intense and widespread, last longer and could result in an increased global aridity (Dai 2011; Prudhomme et al. 2014).

The projected changes in the frequency and intensity of extreme temperatures and drought is expected to result in decreased carbon sequestration by ecosystems and degradation of ecosystems health and loss of resilience (Trumbore et al. 2015). Also affected are many aspects of land functioning and type including agricultural productivity (Lesk et al. 2016a), hydrology (Mosley 2015; Van Loon and Laaha 2015), vegetation productivity and distribution (Xu et al. 2011; Zhou et al. 2014), carbon fluxes and stocks and other biogeochemical cycles (Frank et al. 2015b; Doughty et al. 2015; Schlesinger et al. 2016). Carbon stocks are particularly vulnerable to extreme events due to their large carbon pools and fluxes, potentially large lagged impacts and long recovery times to regain lost stocks (Frank et al. 2015a)(Section 2.2).

2.2.5.3 *Changes in heavy precipitation*

A large number of extreme rainfall events have been documented over the past decades (Coumou and Rahmstorf 2012; Seneviratne et al. 2012a; Trenberth 2012; Westra et al. 2013; Espinoza et al. 2014; Guhathakurta et al. 2017; Taylor et al. 2017; Thompson et al. 2017; Zilli et al. 2017). The observed shift in the trend distribution of precipitation extremes is more distinct than for annual mean precipitation and the global land fraction experiencing more intense precipitation events is larger than expected from internal variability (Fischer and Knutti 2014; Espinoza et al. 2018; Fischer et al. 2013) . As a result of global warming the number of record-breaking rainfall events globally has increased significantly by 12% during the period 1981 to 2010 compared to those expected due to natural multi-decadal climate variability (Lehmann et al. 2015) and the IPCC SR15 reports robust increases in observed precipitation extremes for annual maximum 1-day precipitation (RX1day) and consecutive 5-day precipitation (RX5day) (Hoegh-Guldberg et al. 2018; Schleussner et al. 2017). A number of extreme rainfall events have been attributed to human influence (Min et al. 2011; Pall et al. 2011; Sippel and Otto 2014; Trenberth et al. 2015; Krishnan et al. 2016) and the largest fraction of anthropogenic influence is evident in the most rare and extreme events (Fischer and Knutti 2014).

A warming climate is expected to intensify the hydrological cycle as a warmer climate facilitates more water vapour in the atmosphere, as approximated by the Clausius-Clapeyron (C-C) relationship, with subsequent effects on regional extreme precipitation events (Christensen and Christensen 2003; Pall et al. 2007; Berg et al. 2013; Wu et al. 2013; Guhathakurta et al. 2017; Thompson et al. 2017; Taylor et al. 2017; Zilli et al. 2017)(Manola et al. 2018). Furthermore, changes to the dynamics of the atmosphere amplify or weaken future precipitation extremes at the regional scale (O’Gorman 2015; Pfahl et al. 2017). Continued anthropogenic warming is very likely to increase the frequency and intensity of extreme rainfall in many regions of the globe (Seneviratne et al. 2012a; Mohan and Rajeevan 2017; Prein et al. 2017; Stott et al. 2016) although many GCMs underestimate observed increased trends in heavy precipitation suggesting a substantially stronger intensification of future heavy rainfall than the multi-model mean (Borodina et al. 2017; Min et al. 2011). Furthermore, the response of extreme convective precipitation to warming remains uncertain because GCMs and regional climate models (RCMs) are unable to explicitly simulate sub-grid scale processes such as convection, the hydrological cycle and surface fluxes and have to rely on parameterisation schemes for this (Crétat et al. 2012; Rossow et al. 2013; Wehner 2013; Kooperman et al. 2014; O’Gorman 2015; Larsen et al. 2016; Chawla et al. 2018; Kooperman et al. 2018; Maher et al. 2018; Rowell and Chadwick 2018). High-resolution regional climate models that explicitly resolve convection have a better representation of extreme precipitation but are dependent on the GCM to capture large scale environment in which the extreme event may occur (Ban et al. 2015; Prein et al. 2015; Kendon et al. 2017) . Interannual variability of precipitation extremes in the convective tropics are not well captured by global models (Allan and Liu 2018).

There is low confidence in the detection of long-term observed and projected seasonal and daily trends of extreme snowfall. The narrow rain–snow transition temperature range at which extreme snowfall can occur

1 that is relatively insensitive to climate warming and subsequent large interdecadal variability (Kunkel et al.
2 2013; O’Gorman 2014, 2015).

3 4 **2.2.5.4 Impacts of precipitation extremes on different land cover types**

5 More intense rainfall leads to water redistribution between surface and ground water in catchments as water
6 storage in the soil decreases (green water) and runoff and reservoir inflow increases (blue water) (Liu and
7 Yang 2010; Eekhout et al. 2018). This results in increased surface flooding and soil erosion, increased plant
8 water stress and reduced water security, which in terms of agriculture means an increased dependency on
9 irrigation and reservoir storage (Nainggolan et al. 2012; Favis-Mortlock and Mullen 2011; García-Ruiz et al.
10 2011; Li and Fang 2016; Chagas and Chaffe 2018). As there is high confidence of a positive correlation
11 between global warming and future flood risk, land cover and processes are likely to be negatively impacted,
12 particularly near rivers and in floodplains (Kundzewicz et al. 2014; Alfieri et al. 2016; Winsemius et al.
13 2016; Arnell and Gosling 2016; Alfieri et al. 2017; Wobus et al. 2017).

14
15 In agricultural systems heavy precipitation and inundation can delay planting, increases soil compaction, and
16 causes crop losses through anoxia and root diseases (Posthumus et al. 2009). In tropical regions flooding
17 associated with tropical cyclones can lead to crop failure from both rainfall and storm surge. In some cases
18 flooding can affect yield more than drought, particularly in tropical regions (e.g. India) and in some mid/high
19 latitude regions such as China and central and northern Europe (Zampieri et al. 2017). Waterlogging of
20 croplands and soil erosion also negatively affect farm operations and block important transport routes (Vogel
21 and Meyer 2018; Kundzewicz and Germany 2012). Flooding can be beneficial in drylands if the floodwaters
22 infiltrate and recharge alluvial aquifers along ephemeral river pathways, extending water availability into dry
23 seasons and drought years and support riparian systems and human communities (Kundzewicz and Germany
24 2012; Guan et al. 2015). Globally, the impact of rainfall extremes on agriculture is less than that of
25 temperature extremes and drought, although in some regions and for some crops, extreme precipitation
26 explains a greater component of yield variability, e.g. of maize in the Mid-Western USA and southern Africa
27 (Ray et al. 2015; Lesk et al. 2016b; Vogel et al. 2019) .

28
29 Although many soils on floodplains regularly suffer from inundation, the increases in the magnitude of flood
30 events means that new areas with no recent history of flooding are now becoming severely affected (Yellen
31 et al. 2014). Surface flooding and associated soil saturation often results in decreased soil quality through
32 nutrient loss, reduced plant productivity, stimulates microbial growth and microbial community composition,
33 negatively impacts soil redox and increases greenhouse gas emissions (Bossio and Scow 1998; Niu et al.
34 2014; Barnes et al. 2018; Sánchez-Rodríguez et al. 2019). The impact of flooding on soil quality is
35 influenced by management systems that may mitigate or exacerbate the impact. Although soils tend to
36 recover quickly after floodwater removal, the impact of repeated extreme flood events over longer timescales
37 on soil quality and function is unclear (Sánchez-Rodríguez et al. 2017).

38
39 Flooding in ecosystems may be detrimental through erosion or permanent habitat loss, or beneficial, as a
40 flood pulse brings nutrients to downstream regions (Kundzewicz et al. 2014). Riparian forests can be
41 damaged through flooding; however, increased flooding may also be of benefit to forests where upstream
42 water demand has lowered stream flow, but this is difficult to assess and the effect of flooding on forests is
43 not well studied (Kramer et al. 2008; Pawson et al. 2013). Forests may mitigate flooding, however flood
44 mitigation potential is limited by soil saturation and rainfall intensity (Pilaš et al. 2011; Ellison et al. 2017b).
45 Some grassland species under heavy rainfall and soil saturated conditions responded negatively with
46 decreased reproductive biomass and germination rates (Gellesch et al. 2017), however overall productivity in
47 grasslands remains constant in response to heavy rainfall (Grant et al. 2014).

48
49 Extreme rainfall alters responses of soil CO₂ fluxes and CO₂ uptake by plants within ecosystems and
50 therefore result in changes in ecosystem carbon cycling (Fay et al. 2008; Frank et al. 2015a). Extreme
51 rainfall and flooding limits oxygen in soil which may suppress the activities of soil microbes and plant roots

1 and lower soil respiration and therefore carbon cycling (Knapp et al. 2008; Rich and Watt 2013; Philben et
2 al. 2015). However, the impact of extreme rainfall on carbon fluxes in different biomes differs. For example,
3 extreme rainfall in mesic biomes reduces soil CO₂ flux to the atmosphere and GPP whereas in xeric biomes
4 the opposite is true, largely as a result of increased soil water availability (Knapp and Smith 2001; Heisler
5 and Knapp 2008; Heisler-White et al. 2009; Zeppel et al. 2014; Xu and Wang 2016; Liu et al. 2017b; Connor
6 and Hawkes 2018).

7
8 As shown above greenhouse gas fluxes between the land and atmosphere are affected by climate. The next
9 section assesses these fluxes in greater detail and the potential for land as a carbon sink.

13 **Cross-Chapter Box 3: Fire and Climate Change**

14
15 Raman Sukumar (India), Almut Arneth (Germany), Werner Kurz (Canada), Andrey Sirin (Russian
16 Federation), Louis Verchot (Colombia/The United States of America)

17
18 Fires have been a natural part of Earth's geological past and its biological evolution since at least the late
19 Silurian, about 400 million years ago (Scott 2000). Presently, roughly 3% of the Earth's land surface burns
20 annually which affects both energy and matter exchanges between the land and atmosphere (Stanne et al.
21 2009). Climate is a major determinant of fire regimes through its control of fire weather, as well as through
22 its interaction with vegetation productivity (fuel availability) and structure (fuel distribution and
23 flammability) (Archibald et al. 2013) at the global (Krawchuk and Moritz 2011), regional (Pausas and Paula
24 2012) and local landscape (Mondal and Sukumar 2016) scales. Presently, humans are the main cause of fire
25 ignition with lightning playing a lesser role globally (Bowman et al. 2017; Harris et al. 2016), although the
26 latter factor has been predominantly responsible for large fires in regions such as the North American boreal
27 forests (Veraverbeke et al. 2017). Humans also influence fires by actively extinguishing them, reducing
28 spread and managing fuels.

29 ***Historical trends and drivers in land area burnt***

30
31 While precipitation has been the major influence on wildfire regimes in pre-Industrial times, human
32 activities have become the dominant drivers since then. There was less biomass burning during the 20th
33 century than at any time during the past two millennia as inferred from charcoal sedimentary records (Doerr
34 and Santín 2016), though there has been an increase in the most recent decades (Marlon et al. 2016). Trends
35 in land area burnt have varied regionally (Giglio et al. 2013). Northern Hemisphere Africa has experienced a
36 fire decrease of 1.7 Mha yr⁻¹ (-1.4% yr⁻¹) since 2000, while Southern Hemisphere Africa saw an increase of
37 2.3 Mha yr⁻¹ (+1.8% yr⁻¹) during the same period. Southeast Asia witnessed a small increase of 0.2 Mha yr⁻¹
38 (+2.5% yr⁻¹) since 1997, while Australia experienced a sharp decrease of about 5.5 Mha yr⁻¹ (-10.7% yr⁻¹)
39 during 2001–2011, followed by an upsurge in 2011 that exceeded the annual area burned in the previous 14
40 years. A recent analysis using the Global Fire Emissions Database v.4 (GFED4s) that includes small fires
41 concluded that the net reduction in land area burnt globally during 1998–2015 was -24.3±8.8% (-
42 1.35±0.49% yr⁻¹) (Andela et al. 2017). However, from the point of fire emissions it is important to consider
43 the land cover types which have experienced changes in area burned; in this instance, most of the declines
44 have come from grasslands, savannas and other non-forest land cover types (Andela et al. 2017). Significant
45 increases in forest area burned (with higher fuel consumption per unit area) have been recorded in western
46 and boreal North America (Abatzoglou and Williams 2016; Ansmann et al. 2018) and in boreal Siberia
47 (Ponomarev et al. 2016) in recent times. The 2017 and 2018 fires in British Columbia, Canada, were the
48 largest ever recorded since the 1950s with 1.2 Mha and 1.4 Mha of forest burnt, respectively (Hanes et al.
49 2018) and smoke from these fires reaching the stratosphere over central Europe (Ansmann et al. 2018).

50
51 Climate variability and extreme climatic events such as severe drought, especially those associated with the

1 El Niño Southern Oscillation (ENSO), play a major role in fire upsurges as in equatorial Asia (Huijnen et al.
2 2016). Fire emissions in tropical forests increased by 133% on average during and following six El Niño
3 years compared to six La Niña years during 1997–2016, due to reductions in precipitation and terrestrial
4 water storage (Chen et al. 2017). The expansion of agriculture and deforestation in the humid tropics has also
5 made these regions more vulnerable to drought-driven fires (Davidson et al. 2012; Brando et al. 2014). Even
6 when deforestation rates were overall declining, as in the Brazilian Amazon during 2003–2015, the incidence
7 of fire increased by 36% during the drought of 2015 (Aragão et al. 2018).

9 ***GHG emissions from fires***

10 Emissions from wildfires and biomass burning are a significant source of greenhouse gases (CO₂, CH₄,
11 N₂O), carbon monoxide (CO), carbonaceous aerosols, and an array of other gases including non-methane
12 volatile organic compounds (NMVOC) (Akagi et al. 2011; Van Der Werf et al. 2010). GFED4s has updated
13 fire-related carbon emission estimates biome-wise, regionally and globally, using higher resolution input
14 data gridded at 0.25°, a new burned area dataset with small fires, improved fire emission factors (Akagi et al.
15 2011; Urbanski 2014) and better fire severity characterisation of boreal forests (van der Werf et al. 2017).
16 The estimates for the period 1997–2016 are 2.2 GtC yr⁻¹, being highest in the 1997 El Niño (3.0 GtC yr⁻¹)
17 and lowest in 2013 (1.8 GtC yr⁻¹). Furthermore, fire emissions during 1997–2016 were dominated by
18 savanna (65.3%), followed by tropical forest (15.1%), boreal forest (7.4%), temperate forest (2.3%), peatland
19 (3.7%) and agricultural waste burning (6.3%) (van der Werf et al. 2017).

20
21 Fires not only transfer carbon from land to the atmosphere but also between different terrestrial pools: from
22 live to dead biomass to soil, including partially charred biomass, charcoal and soot constituting 0.12–0.39
23 GtC yr⁻¹ or 0.2–0.6% of annual terrestrial NPP (Doerr and Santín 2016). Carbon from the atmosphere is
24 sequestered back into regrowing vegetation at rates specific to the type of vegetation and other
25 environmental variables (Loehman et al. 2014). Fire emissions are thus not necessarily a net source of carbon
26 into the atmosphere, as post-fire recovery of vegetation can sequester a roughly equivalent amount back into
27 biomass over a time period of one to a few years (in grasslands and agricultural lands) to decades (in forests)
28 (Landry and Matthews 2016). Fires from deforestation (for land use change) and on peatlands (which store
29 more carbon than terrestrial vegetation) obviously are a net source of carbon from the land to the atmosphere
30 (Turetsky et al. 2014); these types of fires were estimated to emit 0.4 GtC yr⁻¹ in recent decades (van der
31 Werf et al. 2017). Peatland fires dominated by smouldering combustion under low temperatures and high
32 moisture conditions can burn for long periods (Turetsky et al. 2014).

34 ***Fires, land degradation/desertification and land-atmosphere exchanges***

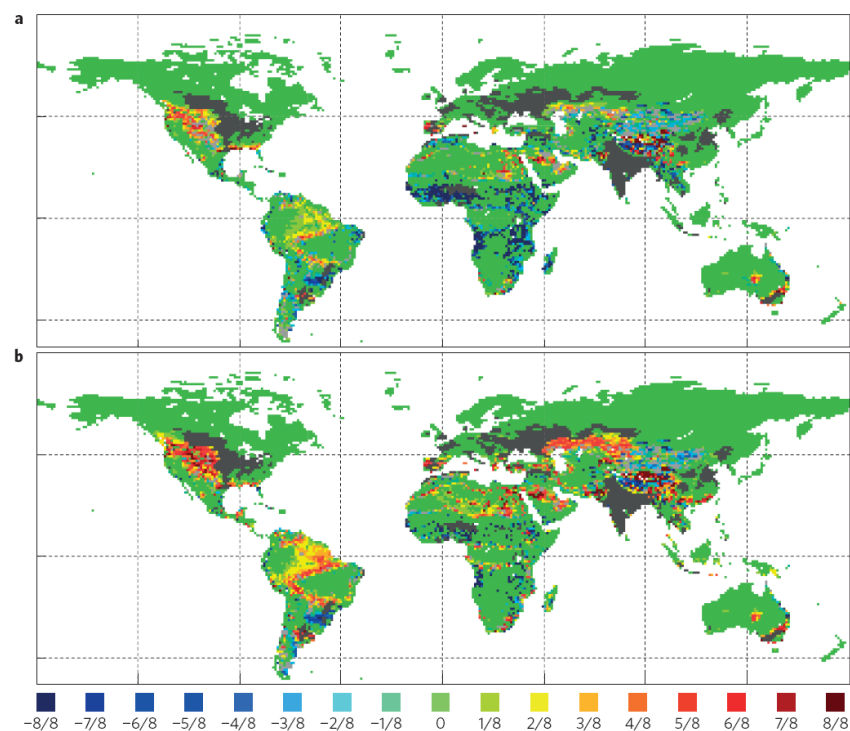
35 Flammable ecosystems are generally adapted to their specific fire regimes (Bond et al. 2005). A fire regime
36 shift alters vegetation and soil properties in complex ways, both in the short- and the long-term, with
37 consequences for carbon stock changes, albedo, fire-atmosphere-vegetation feedbacks and the ultimate
38 biological capacity of the burnt land (Bond et al. 2004; Bremer and Ham 1999; MacDermott et al. 2016;
39 Tepley et al. 2018; Moody et al. 2013; Veraverbeke et al. 2012) A fire-driven shift in vegetation from a
40 forested state to an alternative stable state such as a grassland (Fletcher et al. 2014; Moritz 2015) with much
41 less carbon stock is a distinct possibility. Fires cause soil erosion through action of wind and water (Moody
42 et al. 2013) thus resulting in land degradation (see Chapter 4) and eventually desertification (see Chapter 3).
43 Fires also affect carbon exchange between land and atmosphere through ozone (retards photosynthesis) and
44 aerosol (slightly increases diffuse radiation) emissions; the net effect on global GPP during 2002–2011 is
45 estimated to be -0.86 ± 0.74 GtC yr⁻¹ (Yue and Unger 2018).

47 ***Fires under future climate change***

48 Temperature increase and precipitation decline would be the major driver of fire regimes under future
49 climates as evapotranspiration increases and soil moisture decreases (Pechony and Shindell 2010; Aldersley
50 et al. 2011; Abatzoglou and Williams 2016; Fernandes et al. 2017). The risk of wildfires in future could be
51 expected to change, increasing significantly in North America, South America, central Asia, southern

1 Europe, southern Africa, and Australia (Liu et al. 2010). There is emerging evidence that recent regional
2 surges in wildland fires are being driven by changing weather extremes, thereby signalling geographical
3 shifts in fire proneness (Jolly et al. 2015). Fire weather season has already lengthened by 18.7% globally
4 between 1979 and 2013, with statistically significant increases across 25.3% but decreases only across
5 10.7% of Earth's land surface covered with vegetation; even sharper changes have been observed during the
6 second half of this period (Jolly et al. 2015). Correspondingly, the global area experiencing long fire weather
7 season (defined as experiencing fire weather season greater than one standard deviation (SD) from the mean
8 global value) has increased by 3.1% per annum or 108.1% during 1979–2013. Fire frequencies under 2050
9 conditions are projected to increase by approximately 27% globally, relative to the 2000 levels, with changes
10 in future fire meteorology playing the most important role in enhancing global wildfires, followed by land
11 cover changes, lightning activities and land use, while changes in population density exhibit the opposite
12 effects (Huang et al. 2014).

13
14 However, climate is only one driver of a complex set of environmental, ecological and human factors in
15 influencing fire regimes (Bowman et al. 2011a). While these factors lead to complex projections of future
16 burnt area and fire emissions (Knorr et al. 2016a,b), human exposure to wildland fires could still increase
17 because of population expansion into areas already under high risk of fires (Knorr et al. 2016a,b). There are
18 still major challenges in projecting future fire regimes, and how climate, vegetation and socio/economic
19 factors will interact (Hantson et al. 2016; Harris et al. 2016). There is also need for integrating various fire
20 management strategies, such as fuel-reduction treatments in natural and planted forests, with other
21 environmental and societal considerations to achieve the goals of carbon emissions reductions, maintain
22 water quality, biodiversity conservation and human safety (Moritz et al. 2014; Gharun et al. 2017).



23
24

Cross-Chapter Box 3, Figure 1: The probability of low-fire regions becoming fire prone (positive values), or of fire-prone areas changing to a low-fire state (negative values) between 1971–2000 and 2017–2100 based on eight-Earth system model (ESM) ensembles, two Shared Socio-economic Pathways (SSPs; see (Jiang 2014)) and two Representative Concentration Pathways (RCPs). Light grey: areas where at least one ensemble simulation predicts a positive and one a negative change (lack of agreement). Dark grey: area with >50% past or future cropland. Fire-prone areas are defined as having a fire frequency of >0.01 yr⁻¹ a RCP4.5 emissions with SSP3 demographics. b, RCP8.5 emissions with SSP5 demographics (Knorr et al. 2016a)

In summary, climate change is playing an increasing role in determining wildfire regimes along-side human activity (*medium confidence*), with future climate variability expected to enhance the risk and severity of wildfires in many biomes such as tropical rainforests (*high confidence*). Fire weather seasons have lengthened globally between 1979 and 2013 (*low confidence*). Global land area burned has declined in recent decades, mainly due to less burning in grasslands and savannas (*high confidence*). While drought remains the dominant driver of fire emissions, there has recently been increased fire activity in some tropical and temperate regions during normal to wetter than average years due to warmer temperatures that increase vegetation flammability (*medium confidence*). The boreal zone is also experiencing larger and more frequent fires, and this may increase under a warmer climate (*medium confidence*)

2.3 Greenhouse gas fluxes between land and atmosphere

Land is simultaneously a source and sink for several greenhouse gases (GHGs). Moreover, both natural and anthropogenic processes determine fluxes of GHGs, making it difficult to separate “anthropogenic” and “non-anthropogenic” emissions and removals. A meeting report by the (IPCC 2010a) divided the processes responsible for fluxes from land into three categories: (1) the *direct effects* of anthropogenic activity due to changing land cover and land management; (2) the *indirect effects* of anthropogenic environmental change, such as climate change, carbon dioxide (CO₂) fertilisation, nitrogen deposition; and (3) *natural* climate variability and natural disturbances (e.g. wildfires, windrow, disease). The meeting report (IPCC 2010a) noted that it was impossible with any direct observation to separate direct anthropogenic effects from non-anthropogenic (indirect and natural) effects in the land sector. As a result, different approaches and methods for estimating the anthropogenic fluxes have been developed by different communities to suit their individual purposes, tools and data availability.

The major GHGs exchanged between land and the atmosphere discussed in this chapter are CO₂ (2.3.1), methane (CH₄, Section 2.3.2) and nitrous oxide (N₂O, Section 2.3.3). We estimate the total emissions from Agriculture, Forestry, and Other Land Use (AFOLU) to be responsible for approximately 22% of global anthropogenic GHG emissions over the period 2003-2012 (Smith et al. 2013a; Ciais et al. 2013a) (Table 2.2). The estimate is similar to that reported in AR5 (*high confidence*), with slightly more than half these emissions coming as non-CO₂ GHGs from agriculture. Emissions from AFOLU have remained relatively constant since AR4, although their relative contribution to anthropogenic emissions has decreased due to increases in emissions from the energy sector.

Table 2.2 Summary of average annual land use fluxes aggregated over the decades 2001 to 2010 and 2007 to 2016. We present averages to smooth the effects of inter-annual variability.

Land use emissions	Mt	Gt CO ₂ e ¹	Mt	Gt CO ₂ e
	CH ₄ or N ₂ O		CH ₄ or N ₂ O	
	2001-2010		2007-2016	
Land use CO ₂				
Bookkeeping model average		4.69		4.84
DGVM average		4.75		4.70
FAOSTAT		3.48		2.81
Non-CO ₂ GHGs				

Agricultural CH ₄				
FAOSTAT	130.44	3.65	135.42	3.79
USEPA ²	141.64	3.97	147.83	4.14
EDGAR ³	145.43	4.07	152.68	4.28
Average		3.90		3.97
Agricultural N ₂ O				
FAOSTAT	6.83	1.81	7.34	1.95
USEPA	8.69	2.30	9.32	2.47
EDGAR	5.92	1.57	6.31	1.67
Average		1.89		2.21
Total emissions from land use⁴		10.48		11.02
Total emissions from all sources ⁵		45.29		50.72
Land use emissions: Total emissions		23%		22%

1 Data sources: Bookkeeping models: Hansis et al. 2015; Houghton & Nassikas 2017; DGVM average: Le Quéré et al.
2 2018; FAOSTAT: Tubiello et al. 2013; USEPA: USEPA 2012; EDGAR: Janssens-Maenhout et al. 2017.

3 ¹ All values expressed in units of CO₂eq are based on AR5 100 year Global Warming Potential values without climate-
4 carbon feedbacks (N₂O = 265; CH₄ = 28).

5 ² USEPA data are calculated from country data using IPCC Tier 1 methods (IPCC 2003) at 5-year increments through
6 2005 and are projected estimates from 2010 through 2030

7 ³ EDGAR data are complete only through 2012; 2007-2012 averages were computed in the second data column, for
8 comparison purposes only. They were not used in the calculation of the average fluxes for the 2007-2016 period.

9 ⁴ Total land use emissions were calculated as the sum of the average emissions from the bookkeeping models and the
10 average non-CO₂ GHG data from the different data sources (values in bold).

11 ⁵ Total emissions from all sources were calculated as the sum of total CO₂e emissions values for energy, industrial
12 sources, waste and other emissions from the PRIMAP database (Gütschow et al. 2016). They do not include emissions
13 from international bunkers and shipping.

16 2.3.1 Carbon Dioxide

17
18 This section is divided into four sub-sections (**Figure 2.4**): (1) the total net flux of CO₂ between land and
19 atmosphere, (2) the contributions of AFOLU¹ fluxes and the non-AFOLU land sink to that total net CO₂ flux,
20 (3) the gross emissions and removals comprising the net AFOLU flux, and (4) the gross emissions and
21 removals comprising the land sink. Emissions to the atmosphere are positive; removals from the atmosphere
22 are negative.

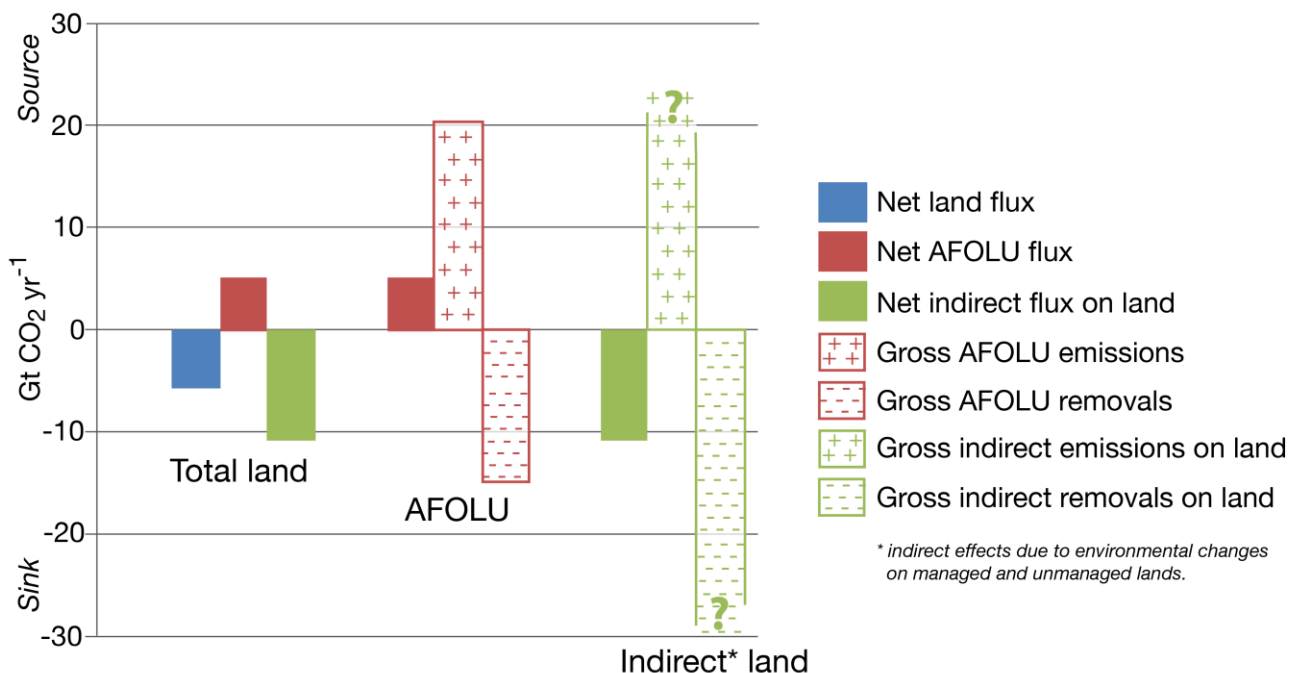
24 2.3.1.1 The total net flux of CO₂ between land and atmosphere

25 The net effects of all anthropogenic and non-anthropogenic processes on managed and unmanaged land
26 result in a net removal of CO₂ from the atmosphere (*high confidence*). This total net land-atmosphere
27 removal (defined here as *the total net land flux*) is estimated to have averaged 6.2 ± 2.6 GtCO₂ yr⁻¹ from
28 2007 to 2016 (**Table 2.3**.) The estimate is determined from summing the AFOLU and non-AFOLU fluxes
29 fluxes due to transient climate change, CO₂ fertilisation, nitrogen deposition calculated by models in the
30 global carbon budget (Le Quéré et al. 2018) and is consistent with inverse modelling techniques based on
31 atmospheric CO₂ concentrations and air transport (range: 5.1–8.8 GtCO₂ yr⁻¹) (Peylin et al. 2013; Van Der
32 Laan-Luijkx et al. 2017; Saeki and Patra 2017; Le Quéré et al. 2018) (See Box 2.2: for methods). A recent
33 inverse analysis, considering carbon transport in rivers and oceans, found a net flux of CO₂ for land within
34 this range, but a lower source from southern lands and a lower sink in northern lands (Resplandy et al. 2018).

35
36 The net removal of CO₂ by land has generally increased over the last 60 years in proportion to total

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1 emissions of CO₂ (*high confidence*). Although land has been a net sink for CO₂ since around the middle of
 2 last century, it was a net source to the atmosphere before that time, primarily as a result of emissions from
 3 AFOLU (Le Quéré et al. 2018).
 4



5
 6
 7 **Figure 2.4 Net and gross fluxes of CO₂ from land (annual averages for 2008-2017).** [Left]: The total net flux
 8 of CO₂ between land and atmosphere (blue) is shown with its two component fluxes: Net AFOLU
 9 emissions (red) and the net land sink (green) due to indirect environmental effects and natural effects on
 10 managed and unmanaged lands. [Middle]: The gross emissions and removals contributing to the net
 11 AFOLU flux. [Right]: The gross emissions and removals contributing to the land sink.

12
 13 **2.3.1.2 Separation of the total net land flux into AFOLU fluxes and the land sink**

14 The total net flux of carbon between land and the atmosphere can be divided into fluxes due to direct human
 15 activities (i.e., AFOLU) and fluxes due to indirect anthropogenic and natural effects (i.e., the land sink)
 16 (Table 2.3). These two components are less certain than their sums, the total net flux of CO₂ between
 17 atmosphere and land. The land sink, estimated with DGVMs, is least certain (Figure 2.5).
 18

19 **Table 2.3 Perturbation of the global carbon cycle caused by anthropogenic activities (GtCO₂ yr⁻¹) (from (Le**
 20 **Quéré et al. 2018)).**

	CO ₂ flux (GtCO ₂ yr ⁻¹), 10-year mean					
	1960–1969	1970–1979	1980–1989	1990–1999	2000–2009	2008–2017
Emissions						
Fossil CO ₂ emissions	11.4 ± 0.7	17.2 ± 0.7	19.8 ± 1.1	23.1 ± 1.1	28.6 ± 1.5	34. ± 1.8
AFOLU net emissions	5.5 ± 2.6	4.4 ± 2.6	4.4 ± 2.6	5.1 ± 2.6	4.8 ± 2.6	5.5 ± 2.6
Partitioning						
Growth in atmosphere	6.2 ± 0.3	10.3 ± 0.3	12.5 ± 0.07	11.4 ± 0.07	14.7 ± 0.07	17.2 ± 0.07
Ocean sink	3.7 ± 1.8	4.8 ± 1.8	6.2 ± 1.8	7.3 ± 1.8	7.7 ± 1.8	8.8 ± 1.8
Land sink	4.4 ± 1.8	7.7 ± 1.5	6.6 ± 2.2	8.8 ± 1.8	9.9 ± 2.6	11.7 ± 2.6

(non-AFOLU)

Budget imbalance	2.2	-1.1	-1.1	0.7	0.7	1.8
Total net land flux (AFOLU - Land sink)	+1.1 ± 3.2	-3.3 ± 3.0	-2.2 ± 3.4	-3.7 ± 2.2	-5.1 ± 3.2	-6.2 ± 3.7

2.3.1.2.1 Fluxes attributed to AFOLU

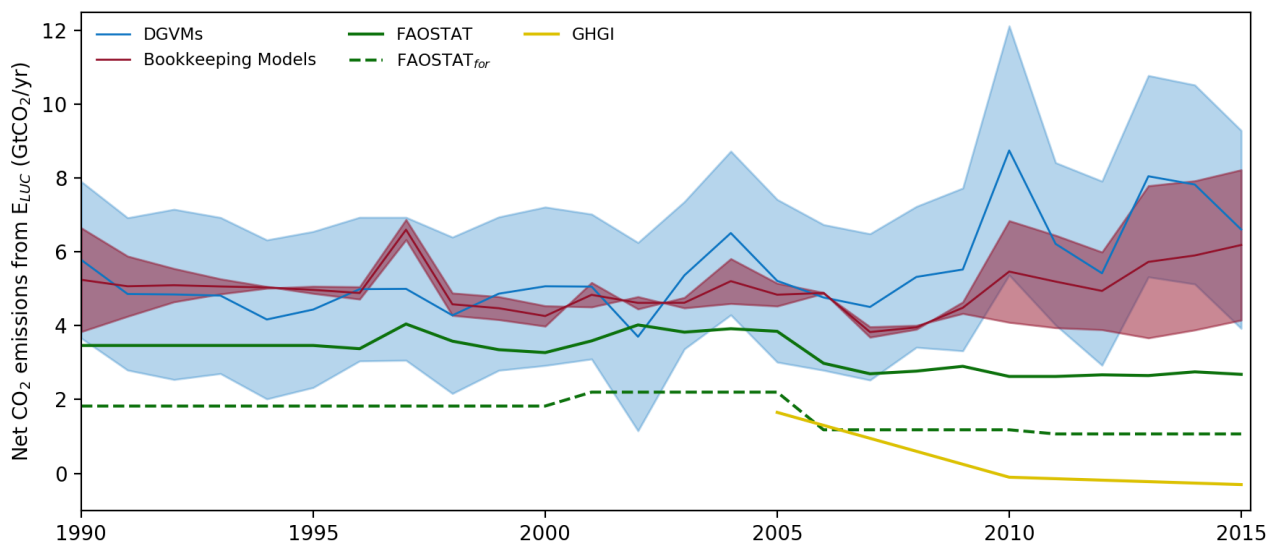
The modelled AFOLU flux was a net emission of 5.5 ± 2.6 GtCO₂ yr⁻¹ for 2008–2017, approximately 14% of total anthropogenic CO₂ emissions (Le Quéré et al. 2018, **Table 2.3**). This net flux was due to direct anthropogenic activities, predominately tropical deforestation, but also afforestation/reforestation, and fluxes due to forest management (e.g. wood harvest) and other types of land management, including agriculture, grasslands and scrub. The AFOLU flux is the mean of two estimates from bookkeeping models (Hansis et al. 2015; Houghton and Nassikas 2017), and this estimated mean is consistent with the mean obtained from an assemblage of Dynamic Global Vegetation Models (DGVMs) (Le Quéré et al. 2018, Box 2.2., **Figure 2.5**), although not all individual DGMVs include the same types of land use. Net CO₂ emissions from AFOLU have been relatively constant since 1900. AFOLU emissions were the dominant anthropogenic emissions until around the middle of the last century when fossil fuel emissions became dominant (Le Quéré et al. 2018). AFOLU activities have resulted in emissions of CO₂ over recent decades (*robust evidence, high agreement*) although there is a wide range of estimates from different methods and approaches (Smith et al. 2014; Houghton et al. 2012; Gasser and Ciais 2013; Pongratz et al. 2014; Tubiello et al. 2015; Grassi et al. 2018) (see Methods Box 2.2., **Figure 2.5** and **Figure 2.7**).

DGVMs and one bookkeeping model (Hansis et al. 2015) used spatially explicit, harmonised land-use change data (LUH2) (Hurt et al. 2017) based on HYDE 3.2. The HYDE data, in turn, are based on changes in the areas of croplands and pastures. In contrast, the Houghton bookkeeping approach (Houghton and Nassikas 2017) used primarily changes in forest area from the FAO Forest Resource Assessment (FAO 2015a) and FAOSTAT to determine changes in land use. To the extent that forests are cleared for land uses other than crops and pastures, estimates from Houghton and Nassikas (2017, 2018) are higher than estimates from DGMVs. In addition, both bookkeeping models (Hansis et al. 2015; Houghton and Nassikas 2017) included estimates of carbon emissions in SE Asia from peat burning from the Global Fire Emissions Database (GFED version 4, (Randerson et al. 2015)) and from peat drainage (Hooijer et al. 2010).

Satellite-based estimates of CO₂ emissions from loss of tropical forests during 2000-2010 corroborate the modelled emissions but are quite variable: 4.8 GtCO₂ yr⁻¹ (Tyukavina et al. 2015), 3.0 GtCO₂ yr⁻¹ (Harris et al. 2015), 3.2 GtCO₂ yr⁻¹ (Achard et al. 2014) and 1.6 GtCO₂ yr⁻¹ (Baccini et al. 2017). Differences in estimates can be explained to a large extent by the different approaches used. For example, the analysis by (Tyukavina et al. 2015) led to a higher estimate because they used a finer spatial resolution. Three of the estimates considered losses in forest area and ignored degradation and regrowth of forests. Baccini et al. (2017) in contrast, included both losses and gains in forest area and losses and gains of carbon within forests (i.e., forest degradation and growth). The four remote sensing studies cited above also reported committed emissions; i.e., all of the carbon lost from deforestation was assumed to be released to the atmosphere in the year of deforestation. In reality, only some of the carbon in trees is not released immediately to the atmosphere at the time of deforestation. The unburned portion is transferred to woody debris and wood products. Both bookkeeping models and DGVMs account for the delayed emissions in growth and decomposition. Finally, the satellite-based estimates do not include changes in soil carbon.

In addition to differences in land-cover data sets between models and satellites, there are many other methodological reasons for differences (See Box 2.2:) (Houghton et al. 2012; Gasser and Ciais 2013; Pongratz et al. 2014; Tubiello et al. 2015). There are different definitions of land-cover type, including forest (e.g. FAO uses a tree cover threshold for forests of 10%; Tyukavina et al. (2017) used 25%), different estimates of biomass and soil carbon density (Mg C ha⁻¹), different approaches to tracking emissions through time (legacy effects), and different types of activity included (e.g. forest harvest, peatland drainage and fires). Most DGVMs only recently (since AR5) included forest management processes, such as tree harvesting and land clearing for shifting cultivation, leading to larger estimates of CO₂ emissions than when

1 these processes are not considered (Arneth et al. 2017; Erb et al. 2018). Grazing management has likewise
 2 been found to have large effects (Sanderman et al. 2017), and is not included in most DGVMs (Pugh et al.
 3 2015; Pongratz et al., 2018).



4
 5 **Figure 2.5** Global net CO₂ emissions due to AFOLU from different approaches (in GtCO₂ yr⁻¹). Red line:
 6 the mean and individual estimates from two bookkeeping models (Houghton and Nassikas 2017; Hansis
 7 et al. 2015). Blue line: the mean from DGVMs run with the same driving data with the pale blue shading
 8 showing the ±1 standard deviation range. Green line: data downloaded from FAOSTAT website
 9 (Tubiello et al. 2013); the dashed line is primarily forest-related emissions, while the solid green line also
 10 includes emissions from peat fires and peat draining. Yellow line: Greenhouse Gas Inventories (GHGI)
 11 based on country reports to UNFCCC (Grassi et al. 2018), data are shown only from 2005 because
 12 reporting in many developing countries became more consistent/reliable after this date. For more details
 13 on methods see Box 2.2:
 14

15 2.3.1.2.2 Nationally reported Greenhouse Gas Inventories (GHGI) values versus global model estimates

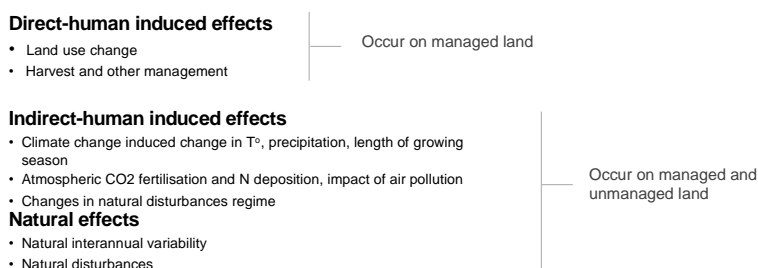
16 There are large differences globally (**Figure 2.5**), between estimates of net anthropogenic land-atmosphere
 17 fluxes of CO₂ from national GHGIs and from global models, and the same is true in many regions (**Figure**
 18 **2.5**). Fluxes reported to the UNFCCC through country GHGIs were noted as about 4.3 GtCO₂ yr⁻¹ lower
 19 (Grassi et al. 2018) than estimates from the bookkeeping model (Houghton et al. 2012a) used in the carbon
 20 budget for AR5 (Conway 2012; Ciais et al. 2013a). The anthropogenic emissions of CO₂ from AFOLU
 21 reported in countries' GHG inventories were 0.1 ± 1.0 GtCO₂ yr⁻¹ globally during 2005 to 2014 (*low*
 22 *confidence*) (Grassi et al. 2018) much lower than emission estimates from the two global bookkeeping
 23 models of 5.1 ± 2.6 GtCO₂ yr⁻¹ over the same time period (Quéré et al. 2018). Transparency and
 24 comparability in estimates can support measuring, reporting and verifying GHG fluxes under the UNFCCC,
 25 and also the global stocktake, which will assess globally the progress towards achieving the long-term goals
 26 of the Paris Agreement. These differences can be reconciled largely by taking account of the different
 27 approaches to defining anthropogenic in terms of different areas of land and treatment of indirect
 28 environmental change (Grassi et al. 2018).
 29

30 To date there has been one study that quantitatively reconciles the global model estimates with GHGIs
 31 (Grassi et al. 2018). The separation of anthropogenic from non-anthropogenic effects is impossible with
 32 direct observation (IPCC 2010a). The different approaches of models and GHGIs to estimating
 33 anthropogenic emissions and removals are shown in (**Figure 2.6**). The difficulty is that *indirect* effects of
 34 environmental changes (e.g. climate change and rising atmospheric CO₂) affect both managed and unmanaged
 35 lands, and some approaches treat these as anthropogenic while others do not. Bookkeeping models (e.g.
 36 Houghton and Nassikas 2017) attempt to estimate the fluxes of CO₂ driven by direct anthropogenic effects
 37 alone. DGVMs model the *indirect* environmental effects of climate and CO₂. If the indirect effects happen
 38 on land experiencing anthropogenic land cover change or management (harvest and regrowth), DGVMs treat
 39 this as anthropogenic. Country GHGIs separately report fluxes due to land conversion (e.g. forests to
 40 croplands) and fluxes due to land management (e.g. forest land remaining forest land). The “managed land
 41 proxy” is used as a pragmatic approach to estimate anthropogenic fluxes on managed lands, whereby

1 countries define the areas they consider managed, and include all of the emission and removals that occur on
 2 those lands. Emissions and removals are caused simultaneously by direct, indirect and natural drivers and are
 3 captured in the reporting, which often relies on inventories.
 4

5 Grassi et al. (2018) demonstrated that estimates of CO₂ emissions from global models and from nationally
 6 reported GHGIs were similar for deforestation and afforestation, but different for managed forests. Countries
 7 generally reported larger areas of managed forests than the models, and the carbon removals by these
 8 managed forests were also larger. The flux due to indirect effects on managed lands was quantified using
 9 post-processing of results from DGVMs, looking at the *indirect* effects of CO₂ and climate change on
 10 secondary forest areas. The derived DGVM *indirect* managed forest flux was found to account for most of
 11 the difference between the bookkeeping models and the inventories.
 12

a) Effects of various factors on the forest CO₂ fluxes and where they occur



b) Conceptual differences in defining the anthropogenic land CO₂ flux

IPCC AR5 and Global Carbon Budget:

Bookkeeping models
 "Land Use Change"

	Managed land	Unmanaged land
Direct human induced effects	✓	
Indirect human induced effects		
Natural effects		

DGVMs:
 "Land Use Change" and "Land Sink"

	Managed land	Unmanaged land
Direct human induced effects	✓	
Indirect human induced effects	✓	✓
Natural effects		

Country GHG inventories:

"AFOLU (LULUCF)"

	Managed land	Unmanaged land
Direct human induced effects	✓	
Indirect human induced effects	✓	
Natural effects	✓	

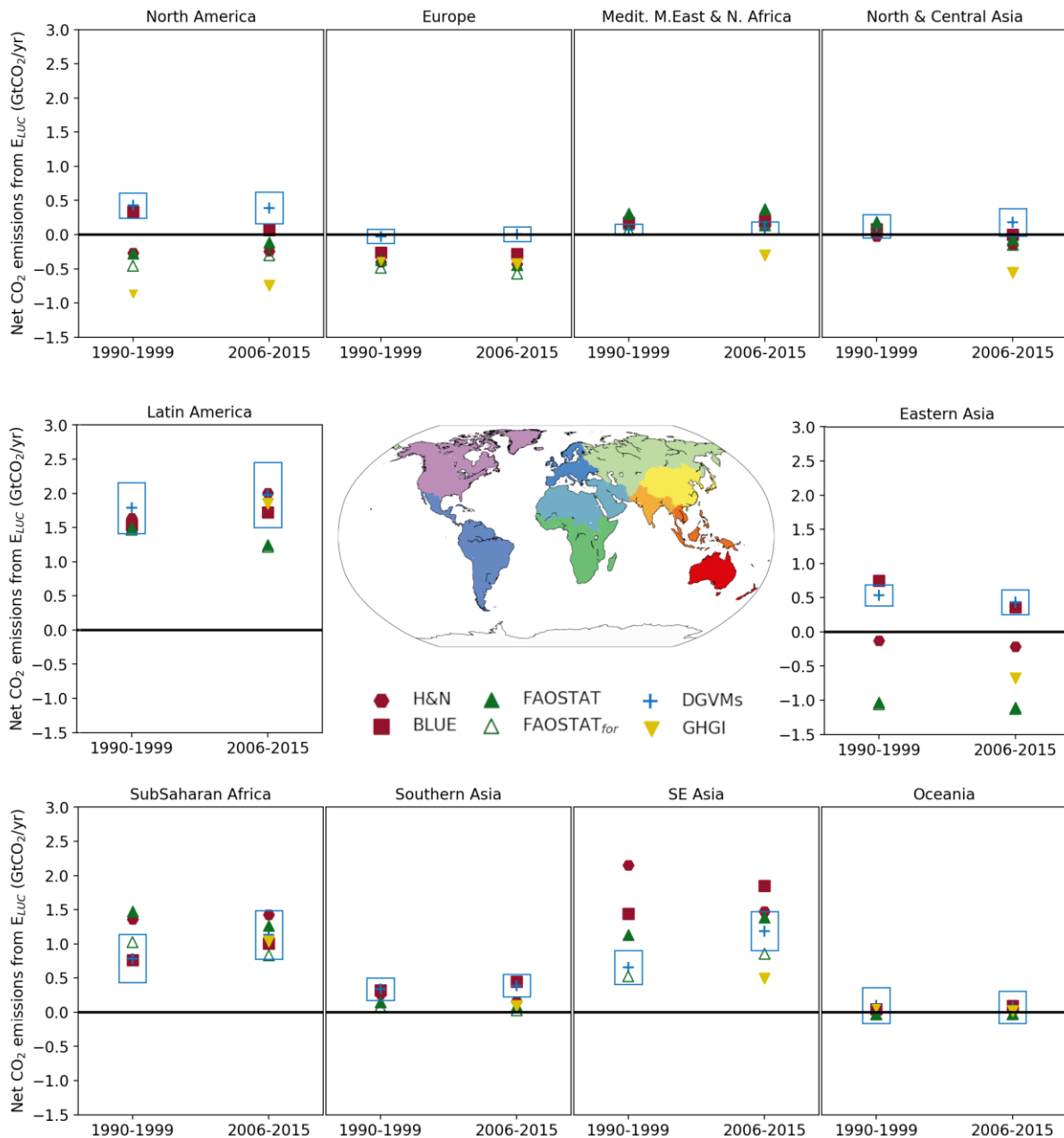
13
 14
 15 **Figure 2.6 Summary of the main conceptual differences between GHG Inventories and global models in**
 16 **considering what is the "anthropogenic land CO₂ flux". Adapted from Grassi et al. (2018): a) Effects of**
 17 **key processes on the land flux as defined by IPCC (2010) including where these effects occur (in**
 18 **managed and/or unmanaged lands); How these effects are captured in '(a) bookkeeping models that do**
 19 **not explicitly model the effects of environmental change, although some is implicitly captured in data on**
 20 **carbon densities and growth and decay rates; (b) Dynamic Global Vegetation Models (DGVMs) that**
 21 **include the effects of environmental change on all lands, and run the models with and without land use**
 22 **change to diagnose "land use change", the "land sink" is then conceptually assumed to be a natural**
 23 **response of land to the anthropogenic perturbation of environmental change, models include the effects**
 24 **of inter-annual climate variability, and some include fires but no other natural disturbances (c) GHG**
 25 **Inventories reported by countries to the UNFCCC that report all fluxes in areas the countries define as**
 26 **"managed land" but do not report unmanaged land. This is the CO₂ flux due to Land Use Land Use**
 27 **Change and Forestry (LULUCF) which is a part of the overall AFOLU (Agriculture, Forestry and**
 28 **Other Land Use) flux. The area of land considered as managed in the inventories is greater than that**
 29 **considered as subject to direct management activities (harvest and regrowth) in the models.**
 30

31 **2.3.1.2.3 Regional differences**

32 Figure 2.7 shows regional differences in emissions due to AFOLU. Recent increases in deforestation rates in
 33 some tropical countries have been partially balanced by increases in forest area in India, China, the USA and

1 Europe (FAO-FRA 2015). The trend in emissions from AFOLU since the 1990s is *uncertain* because some
 2 data suggest a declining rate of deforestation (FAO-FRA 2015), while data from satellites suggest an
 3 increasing rate (Kim 2014; Hansen et al. 2012). The disagreement results in part from differences in the
 4 definition of forest and approaches to estimating deforestation. The FAO defines deforestation as the
 5 conversion of forest to another land use (FAO-FRA 2015), while the measurement of forest loss by satellite
 6 may include wood harvests (forests remaining forests) and natural disturbances that are not directly caused
 7 by anthropogenic activity (e.g., forest mortality from droughts and fires). Trends in anthropogenic and
 8 natural disturbances may be in opposite directions. For example, recent drought-induced fires in the Amazon
 9 have increased the emissions from wildfires at the same time that emissions from anthropogenic
 10 deforestation have declined (Aragão et al. 2018). Furthermore, there have been advances since AR5 in
 11 estimating the GHG effects of different types of forest management (e.g. (Valade et al. 2017). Overall, there
 12 is *robust evidence and high agreement* for a net loss of forest area and tree cover in the tropics and a net
 13 gain, mainly of secondary forests and sustainably managed forests, in the temperate and boreal zones
 14 (Chapter 1).

15



16
 17
 18

Figure 2.7 Regional trends in net anthropogenic land-atmosphere CO₂ flux from a range of different approaches (in GtCO₂ yr⁻¹). Red symbols: bookkeeping models (hexagon - Houghton and Nassikas 2017;

1 square - Hansis et al. 2015). Blue cross: the mean from DGVMs with the box showing the 1 standard
 2 deviation range. Green triangles: downloaded from FAOSTAT website; the open triangle is primarily
 3 forest-related emissions, while the closed triangle includes emission from peat fires and peat drainage.
 4 Yellow inverted triangle: Greenhouse Gas Inventories (GHGI) LULUCF flux (Land Use Land Use
 5 Change and Forestry – part of AFOLU) based on country reports to UNFCCC (Grassi et al. 2018) –
 6 data for developing countries is only shown for 2006-2015 because reporting in many developing
 7 countries became more consistent/reliable after 2005. For more details on methods see Box 2.2:.

9 2.3.1.2.4 Processes responsible for the land sink

10 Just over half of total net anthropogenic CO₂ emissions (AFOLU and fossil fuels) were taken up by oceanic
 11 and land sinks (Table 2.3) (*robust evidence, high agreement*). The land sink was referred to in AR5 as the
 12 “residual terrestrial flux,” as it was not estimated directly, but calculated by difference from the other directly
 13 estimated fluxes in the budget (Table 2.3). In the 2018 budget (Le Quéré et al. 2018), the land sink term was
 14 instead estimated directly by DGVMs, leaving a budget imbalance of 2.2 GtCO₂ yr⁻¹ (sources overestimated
 15 or sinks underestimated). The budget imbalance may result from variations in oceanic uptake, or from
 16 uncertainties in fossil fuel or AFOLU emissions, as well as from land processes not included in DGVMs.

17
 18 The land sink is thought to be driven largely by the indirect effects of environmental change (e.g., climate
 19 change, increased atmospheric CO₂ concentration nitrogen deposition) on unmanaged and managed lands
 20 (*robust evidence, high agreement*). The land sink has generally increased since 1900 and was a net sink of
 21 11.7 ± 3.7 GtCO₂ yr⁻¹ during the period 2008 to 2017 (Table 2.3), absorbing 29% of global anthropogenic
 22 emissions of CO₂. The land sink has slowed the rise in global land-surface air temperature by 0.09 ± 0.02°C
 23 since 1982 (*medium confidence*) (Zeng et al. 2017).

24
 25 The rate of CO₂ removal by land accelerated from -0.026 ± 0.24 GtCO₂ yr⁻¹ during the warming period (1982
 26 to 1998) to -0.436 ± 0.260 GtCO₂ yr⁻¹ during the warming hiatus (1998-2012). One explanation is that
 27 respiration rates were lower during the warming hiatus (Ballantyne et al. 2017). However, the lower rate of
 28 growth in atmospheric CO₂ during the warming hiatus may have resulted, not from lower rates of respiration,
 29 but from declining emissions from AFOLU (lower rates of tropical deforestation and increased forest growth
 30 in northern mid-latitudes (Piao et al. 2018). Changes in the growth rate of atmospheric CO₂, by themselves,
 31 do not identify the processes responsible, and the cause of the variation is uncertain.

32
 33 While year-to-year variability in the indirect land sink is high in response to climate variability, DGVM
 34 fluxes are far more influenced on decadal time scales by CO₂ fertilisation. A DGVM intercomparison (Sitch
 35 et al. 2015) for 1990 to 2009 found that CO₂ fertilisation alone contributed a mean global removal of -10.54
 36 ± 3.68 GtCO₂ yr⁻¹ (trend -0.444 ± 0.202 GtCO₂ yr⁻¹). Data from forest inventories around the world
 37 corroborate the modelled land sink (Pan et al. 2011a). The geographic distribution of the non-AFOLU land
 38 sink is less certain. While it seems to be distributed globally, its distribution between the tropics and non-
 39 tropics is estimated to be between 1:1 (Pan et al. 2011a) and 1:2 (Houghton et al. 2018).

40
 41 As described in Box 2.3, rising CO₂ concentrations have a fertilising effect on land, while climate has mixed
 42 effects; e.g., rising temperature increases respiration rates and may enhance or reduce photosynthesis
 43 depending on location and season, while longer growing seasons might allow for higher carbon uptake.
 44 However, these processes are not included in DGVMs, which may account for at least some of the land sink.
 45 For example, a decline in the global area burned by fires each year (Andela et al. 2017) accounts for an
 46 estimated net sink (and/or reduced emissions) of 0.5 GtCO₂ yr⁻¹ (Arora and Melton 2018) (*limited evidence,*
 47 *medium agreement*) (boreal forests represent an exception to this decline (Kelly et al. 2013)). The reduction
 48 in burning not only reduces emissions; it also allows more growth of recovering forests. There is also an
 49 estimated net carbon sink of about the same magnitude (0.5 GtCO₂ yr⁻¹) as a result of soil erosion from
 50 agricultural lands and redeposition in anaerobic environments where respiration is reduced (Wang et al.
 51 2017d) (*limited evidence, low agreement*). A recent study attributes an increase in land carbon to a longer-
 52 term (1860-2005) aerosol-induced cooling (Zhang et al. 2019). Recent evidence also suggests that DGVMs
 53 and Earth System Models underestimate the effects of drought on CO₂ emissions (Humphrey et al. 2018;
 54 Green et al. 2019; Kolus et al. 2019).

55 2.3.1.3 Gross emissions and removals contributing to AFOLU emissions

56 The modelled AFOLU flux of 5.5 ± 3.7 GtCO₂ yr⁻¹ over the period 2008 to 2017 represents a net value. It
 57

1 consists of both gross emissions of CO₂ from deforestation, forest degradation, and the oxidation of wood
2 products, as well as gross removals of CO₂ in forests and soils recovering from harvests and agricultural
3 abandonment (Figure 2.4). The uncertainty of these gross fluxes is high because few studies report gross
4 fluxes from AFOLU. Houghton and Nassikas (2017) estimated gross emissions to be as high as 20.2 GtCO₂
5 yr⁻¹ (*limited evidence, low agreement*) (Figure 2.4), and even this may be an underestimate because the land-
6 use change data used from FAOSTAT (Tubiello et al. 2013) is itself a net of all changes within a country.
7

8 Gross emissions and removals of CO₂ result from rotational uses of land, such as wood harvest and shifting
9 cultivation, including regrowth. These gross fluxes are more informative for assessing the timing and
10 potential for mitigation than estimates of net fluxes, because the gross fluxes include a more complete
11 accounting of individual activities. Gross emissions from rotational land use in the tropics are approximately
12 37% of total CO₂ emissions, rather than 14%, as suggested by net AFOLU emissions (Houghton and
13 Nassikas 2018). Further, if the forest is replanted or allowed to regrow, gross removals of nearly the same
14 magnitude would be expected to continue for decades.
15

16 **2.3.1.4 Gross emissions and removals contributing to the non-anthropogenic land sink**

17 The *net* land sink averaged 11.7 GtCO₂ yr⁻¹ over 2008-2017 (*robust evidence, medium agreement*) (Table
18 2.3.2), but its gross components have not been estimated at the global level. There are many studies that
19 suggest increasing emissions of carbon due to indirect environmental effects and natural disturbance, for
20 example temperature-induced increases in respiration rates (Bond-Lamberty et al. 2018); increased tree
21 mortality (Brienen et al. 2015; Berdanier and Clark 2016; McDowell et al. 2018); and thawing permafrost
22 (Schuur et al. 2015). The global carbon budget indicates that land and ocean sinks have *increased* over the
23 last six decades in proportion to total CO₂ emissions (Le Quéré et al. 2018) (*robust evidence, high*
24 *agreement*). That means that any emissions must have been balanced by even larger removals (likely driven
25 by CO₂ fertilisation, climate change, nitrogen deposition, erosion and redeposition of soil carbon, a reduction
26 in areas burned, aerosol-induced cooling, and changes in natural disturbances, Box 2.3)
27

28 Climate change is expected to impact terrestrial biogeochemical cycles via an array of complex feedback
29 mechanisms that will act to either enhance or decrease future CO₂ emissions from land. Because the gross
30 emissions and removals from environmental changes are not constrained at present, the balance of future
31 positive and negative feedbacks remains uncertain. Estimates from climate models included in AR5, CMIP5
32 (Coupled Model Intercomparison Project, 5), exhibit large differences for different carbon and nitrogen cycle
33 feedbacks and how they change in a warming climate (Anav et al. 2013; Friedlingstein et al. 2006;
34 Friedlingstein, et al. 2014) . The differences are in large part due to the uncertainty regarding how primary
35 productivity and soil respiration will respond to environmental changes, with many of the models not even
36 agreeing on the sign of change. Furthermore, many models do not include a nitrogen cycle, which may limit
37 the CO₂ fertilisation effect in the future (see Box 2.3). There is an increasing amount of observational data
38 available and methods to constrain models (e.g. Cox et al. 2013; Prentice, et al., 2015) which can reduce
39 uncertainty.
40

41 **2.3.1.5 Potential impact of mitigation on atmospheric CO₂ concentrations**

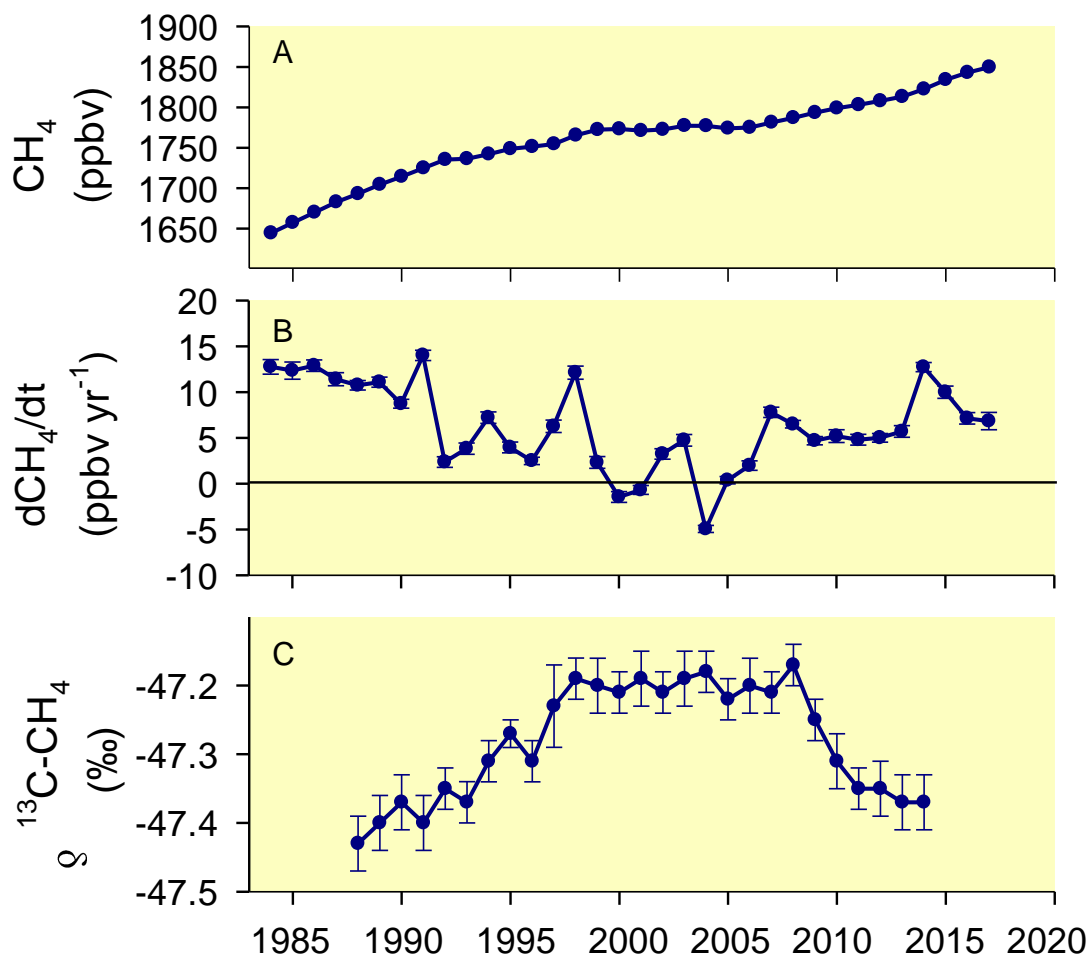
42 If CO₂ concentrations decline in the future as a result of low emissions and large negative emissions, the
43 global land and ocean sinks are expected to weaken (or even reverse). The oceans are expected to release
44 CO₂ back to the atmosphere when the concentration declines (Ciais et al. 2013a; Jones et al. 2016). This
45 means that to maintain atmospheric CO₂ and temperature at low levels, both the excess CO₂ from the
46 atmosphere and the CO₂ progressively outgassed from the ocean and land sinks will need to be removed.
47 This outgassing from the land and ocean sinks is called the “rebound effect” of the global carbon cycle (Ciais
48 et al. 2013a). It will reduce the effectiveness of negative emissions and increase the deployment level needed
49 to achieve a climate stabilisation target (Jackson et al. 2017; Jones et al. 2016) (*limited evidence, high*
50 *agreement*).
51

52 **2.3.2 Methane**

53 **2.3.2.1 Atmospheric trends**

54 In 2017, the globally averaged atmospheric concentration of CH₄ was 1850 ± 1 ppbv (Figure 2.8A).
55 Systematic measurements of atmospheric CH₄ concentrations began in the mid-1980s and trends show a
56

1 steady increase between the mid-1980s and early-1990s, slower growth thereafter until 1999, a period of no
 2 growth between 1999 and 2006, followed by a resumption of growth in 2007. The growth rates show very
 3 high inter-annual variability with a negative trend from the beginning of the measurement period until about
 4 2006, followed by a rapid recovery and continued high inter-annual variability through 2017 (Figure 2.8B).
 5 The growth rate has been higher over the past 4 years (*high confidence*) (Nisbet et al. 2019). The trend in
 6 $\delta^{13}\text{C}-\text{CH}_4$ prior to 2000 with less depleted ratios indicated that the increase in atmospheric concentrations
 7 was due to thermogenic (fossil) CH_4 emissions; the reversal of this trend after 2007 indicates a shift to
 8 biogenic sources (Figure 2.8C).



9
10
11 **Figure 2.8 Globally averaged atmospheric CH_4 mixing ratios (Frame A) and instantaneous rates of change**
 12 **(Frame B) and C isotope /variation (Frame C) Data sources: NOAA/ESRL**
 13 **(www.esrl.noaa.gov/gmd/ccgg/trends_ch4/)(Dlugokencky et al. 1994) and Schaefer et al. (2016).**
 14

15 Understanding the underlying causes of temporal variation in atmospheric CH_4 concentrations is an active
 16 area of research. Several studies concluded that inter-annual variability of CH_4 growth was driven by
 17 variations in natural emissions from wetlands (Rice et al. 2016; Bousquet et al. 2006; Bousquet et al. 2011;
 18 Bousquet et al. 2011b). These modelling efforts concluded that tropical wetlands were responsible for
 19 between 50 and 100% of the inter-annual fluctuations and the renewed growth in atmospheric concentrations
 20 after 2007. However, results were inconsistent for the magnitude and geographic distribution of the wetland
 21 sources between the models. Pison et al. (2013) used two atmospheric inversion models and the ORCHIDEE
 22 model and found greater uncertainty in the role of wetlands in inter-annual variability between 1990 and
 23 2009 and during the 1999-2006 pause. Poulter et al. (2017) used several of biogeochemical models and
 24 inventory-based wetland area data to show that wetland CH_4 emissions increases in the boreal zone have
 25 been offset by decreases in the tropics and concluded that wetlands have not contributed significantly to
 26 renewed atmospheric CH_4 growth.
 27

1 The models cited above assumed that atmospheric hydroxyl radical (OH) sink over the period analysed did
2 not vary. OH reacts with CH₄ as the first step toward oxidation to CO₂. In global CH₄ budgets, the
3 atmospheric OH sink has been difficult to quantify because its short lifetime (~1 second) and its distribution
4 is controlled by precursor species that have non-linear interactions (Taraborrelli et al., 2012; Prather et al.,
5 2017). Understanding of the atmospheric OH sink has evolved recently. The development of credible time
6 series of methyl chloroform (MCF: CH₃CCl₃) observations offered a way understand temporal dynamics of
7 OH abundance and applying this to global budgets further weakened the argument for the role of wetlands in
8 determining temporal trends since 1990. Several authors used the MCF approach and concluded that changes
9 in the atmospheric OH sink explained a large portion of the suppression in global CH₄ concentrations
10 relative to the pre-1999 trend (Turner et al. 2017; Rigby et al. 2013; McNorton et al. 2016). These studies
11 could not reject the null hypothesis that OH has remained constant in recent decades and they did not suggest
12 a mechanism for the inferred OH concentration changes (Nisbet et al. 2019). Nicely et al. (2018) used a
13 mechanistic approach and demonstrated that variation in atmospheric OH was much lower than what MCF
14 studies and found that positive trends in OH due to the effects of water vapour, nitrogen oxides (NO_x),
15 tropospheric ozone, and expansion of the tropical Hadley cells offsets the decrease in OH that is expected
16 from increasing atmospheric CH₄ concentrations.

17
18 The depletion of δ¹³C_{atm} beginning in 2009 could be due to changes in several sources. Decreased fire
19 emissions combined with increased tropical wetland emissions compared to earlier years could explain the
20 δ¹³C perturbations to atmospheric CH₄ sources (Worden et al. 2017; Schaefer et al. 2016). However, because
21 tropical wetland emissions are higher in the Southern Hemisphere, and the remote sensing observations show
22 that CH₄ emissions increases are largely in the north tropics (Bergamaschi et al. 2013; Melton et al. 2013;
23 Houweling et al. 2014), an increased wetland source does not fit well with the southern hemisphere δ¹³C
24 observations. New evidence shows that tropical wetland CH₄ emissions are significantly underestimated,
25 perhaps by a factor of 2, because estimates do not account for release by tree stems (Pangala et al. 2017).
26 Several authors have concluded that agriculture is a more probable source of increased emissions, and
27 particularly from rice and livestock in the tropics, which is consistent with inventory data (Wolf et al. 2017;
28 Patra et al. 2016; Schaefer et al. 2016).

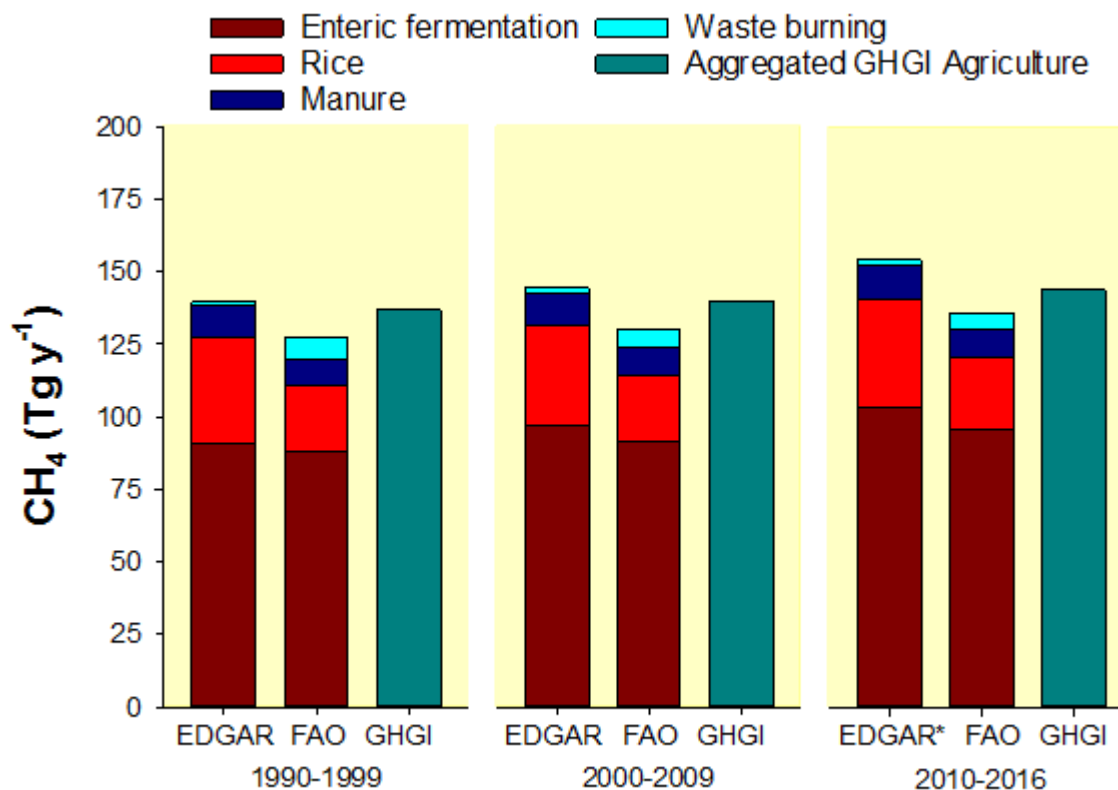
29
30 The importance of fugitive emissions in the global atmospheric accumulation rate is growing (*medium*
31 *evidence, high agreement*). The increased production of natural gas in the US from the mid 2000's is of
32 particular interest because it coincides with renewed atmospheric CH₄ growth (Rice et al. 2016; Hausmann et
33 al. 2015). Reconciling increased fugitive emissions with increased isotopic depletion of atmospheric CH₄
34 indicates that there are *likely* multiple changes in emissions and sinks that affect atmospheric accumulation
35 (*medium confidence*).

36
37 With respect to atmospheric CH₄ growth rates, we conclude that there is significant and ongoing
38 accumulation of CH₄ in the atmosphere (*very high confidence*). The reason for the pause in growth rates and
39 subsequent renewed growth is at least partially associated with land use and land use change. Evidence that
40 variation in the atmospheric OH sink plays a role in the year to year variation of the CH₄ is accumulating,
41 but results are contradictory (*medium evidence, low agreement*) and refining this evidence is constrained by
42 lack of long-term isotopic measurements at remote sites, particularly in the tropics. Fugitive emissions *likely*
43 contribute to the renewed growth after 2006 (*medium evidence, high agreement*). Additionally, the recent
44 depletion trend of ¹³C isotope in the atmosphere indicates that growth in biogenic sources explains part of the
45 current growth and that biogenic sources make up a larger proportion of the source mix compared to the
46 period before 1997 (*robust evidence, high agreement*). In agreement with the findings of AR5, we conclude
47 that wetlands are important drivers of inter-annual variability and current growth rates (*medium evidence,*
48 *high agreement*). Ruminants and the expansion of rice cultivation are also important contributors to the
49 current growth trend (*medium evidence, high agreement*).

50 51 2.3.2.2 Land use effects

52 Agricultural emissions are predominantly from enteric fermentation and rice, with manure management and
53 waste burning contributing small amounts (Figure 2.9). Since 2000, livestock production has been
54 responsible for 33% of total global emissions and 66% of agricultural emissions (Source: EDGAR 4.3.2
55 database, accessed May 2018, (USEPA 2012; Tubiello et al. 2014; Janssens-Maenhout et al. 2017b). Asia
56 has the largest livestock emissions (37%) and emissions in the region have been growing by around 2% per

1 year over the same period. North America is responsible for 26% and emissions are stable; Europe is
 2 responsible for around 8% of emissions, and these are decreasing slightly (<1% per year). Africa is
 3 responsible for 14%, but emissions are growing fastest in this region at around 2.5% y^{-1} . In Latin America
 4 and the Caribbean, livestock emissions are decreasing at around 1.6% per year and the region makes up 16%
 5 of emissions. Rice emissions are responsible for about 24% of agricultural emissions, and 89% of these are
 6 from Asia. Rice emissions are increasing by 0.9% per year in that region. These trends are predicted to
 7 continue through 2030 (USEPA 2013).



8 **Figure 2.9 Average agricultural CH₄ emissions estimates from 1990. Sub-sectorial agricultural emissions are**
 9 **based on the Emissions Database for Global Atmospheric Research (EDGAR v4.3.2; Janssens-**
 10 **Maenhout et al. 2017a); FAOSTAT (Tubiello et al. 2013); and National GHGI data (Grassi et al. 2018).**
 11 **GHGI data are aggregate values for the sector.**

12 * Note that EDGAR data are complete only through 2012; the data in the right-hand panel represent the
 13 three years 2010-2012 and are presented for comparison.
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Upland soils are a net sink of atmospheric CH₄, but soils both produce and consume the gas. On the global
 scale climatic zone, soil texture, and land cover have an important effect on CH₄ uptake in upland soils (Tate
 2015; Yu et al. 2017; Dutaur and Verchot 2007). Boreal soils take up less than temperate or tropical soils,
 coarse textured soils take up more CH₄ than medium and fine textured soils, and forests take up more than
 other ecosystems. Low levels of nitrogen fertilisation or atmospheric deposition can affect the soil microbial
 community and stimulate soil CH₄ uptake in nitrogen limited soils, while higher fertilisation rates decrease
 uptake (Edwards et al. 2018; Zhuang et al., 2013). Globally, N fertilisation on agricultural lands may have
 suppressed CH₄ oxidation by as much as 26 Tg between 1998 and 2004 (Zhuang et al., 2013)(*low confidence* ,
low agreement). The effect of N additions is cumulative and repeated fertilisation events have progressively
 greater suppression effects (*robust evidence, high agreement*) (Tate 2015). Other factors like higher
 temperatures, increased atmospheric concentrations, and changes in rainfall patterns stimulate soil CH₄
 consumption in unfertilised ecosystems. Several studies (Yu et al. 2017; Xu et al. 2016; Curry 2009) have
 shown that globally, uptake has been increasing during the second half of the 20th century and it is expected
 to continue to increase by as much as 1 Tg in the 21st century, particularly in forests and grasslands (*medium*
evidence, high agreement).

Northern peatlands (40°-70°N) are a significant source of atmospheric CH₄, emitting about 48 Tg CH₄, or

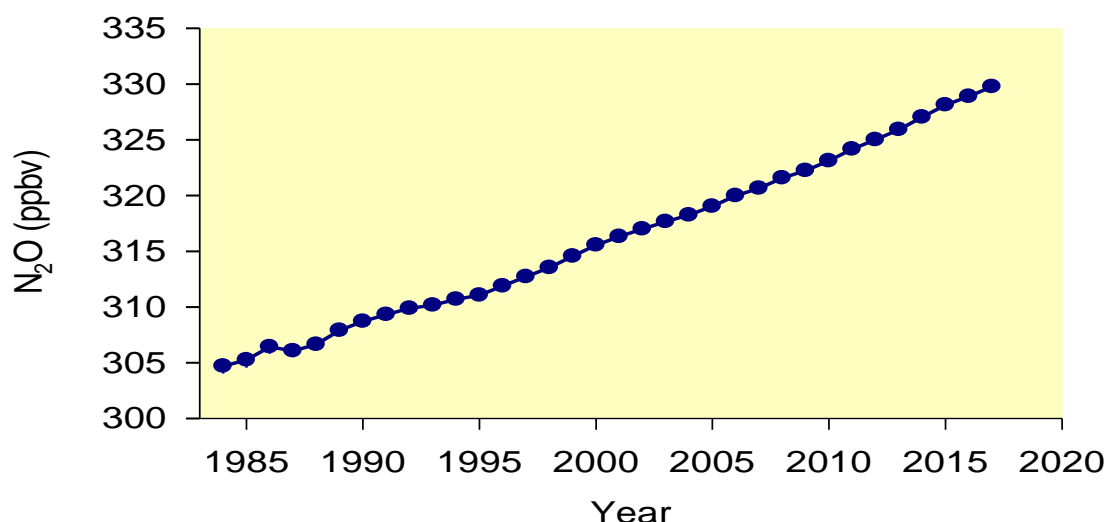
1 about 10% of the total emissions to the atmosphere (Zhuang et al. 2006; Wuebbles and Hayhoe 2002). CH₄
 2 emissions from natural northern peatlands are highly variable with the highest rate from fens (*medium*
 3 *evidence, high agreement*). Peatland management and restoration, alters the exchange of CH₄ with the
 4 atmosphere (*medium evidence, high agreement*). Management of peat soils typically converts them from CH₄
 5 sources to sinks (Augustin et al. 2011; Strack and Waddington 2008; Abdalla et al. 2016) (*robust evidence,*
 6 *high agreement*). While restoration decreases CO₂ emissions (see Section 4.9.4), CH₄ emissions often
 7 increase relative to the drained conditions (*robust evidence, high agreement*) (Osterloh et al. 2018; Christen
 8 et al. 2016; Koskinen et al. 2016; Tuittila et al. 2000; Vanselow-Algan et al. 2015; Abdalla et al. 2016).
 9 Drained peatlands are usually considered to be negligible methane sources, but they emit CH₄ under wet
 10 weather conditions and from drainage ditches (Drösler et al. 2013; Sirin et al. 2012). While ditches cover
 11 only a small percentage of the drained area, emissions can be sufficiently high that drained peatlands emit
 12 comparable CH₄ as undrained ones (*medium evidence, medium agreement*) (Sirin et al. 2012; Wilson et al.
 13 2016).

14
 15 Because of the large uncertainty in the tropical peatland area, estimates of the global flux are highly
 16 uncertain. A meta-analysis of the effect of conversion of primary forest to rice production showed that
 17 emissions increased by a factor of 4 (*limited evidence, high agreement*) (Hergoualc'h and Verchot, 2012).
 18 For land uses that required drainage, emissions decreased by a factor of 3 (*limited evidence, high*
 19 *agreement*). There are no representative measurements of emissions from drainage ditches in tropical
 20 peatlands.

21 2.3.3 Nitrous Oxide

22 2.3.3.1 Atmospheric trends

23
 24 The atmospheric abundance of N₂O has increased since 1750, from a pre-industrial concentration of 270
 25 ppbv to 330 ppbv in 2017 (U.S. National Oceanographic and Atmospheric Agency, Earth Systems Research
 26 Laboratory; Figure 2.10) (*high agreement, robust evidence*). The rate of increase has also increased, from
 27 approximately 0.15 ppbv yr⁻¹ 100 years ago, to 0.85 ppbv yr⁻¹ over the period 2001 to 2015 (Wells et al.
 28 2018). Atmospheric N₂O isotopic composition (^{14/15}N) was relatively constant during the pre-industrial
 29 period (Prokopiou et al. 2018) and shows a decrease in the δ¹⁵N as the N₂O mixing ratio in the atmosphere
 30 has increased between 1940 and 2005. This recent decrease indicates as that terrestrial sources are the
 31 primary driver of increasing trends and marine sources contribute around 25% (Snider et al. 2015).
 32 Microbial denitrification and nitrification processes are responsible for more than 80% of total global N₂O
 33 emissions, which includes natural soils, agriculture, and oceans, with the remainder coming from non-
 34 biological sources such as biomass burning and fossil-fuel combustion (Fowler et al. 2015). The isotopic
 35 trend also indicates a shift from denitrification to nitrification as the primary source of N₂O as a result of the
 36 use of synthetic nitrogen (N) fertiliser (*high evidence, high agreement*) (Park et al. 2012; Toyoda et al. 2013;
 37 Snider et al. 2015; Prokopiou et al. 2018).
 38

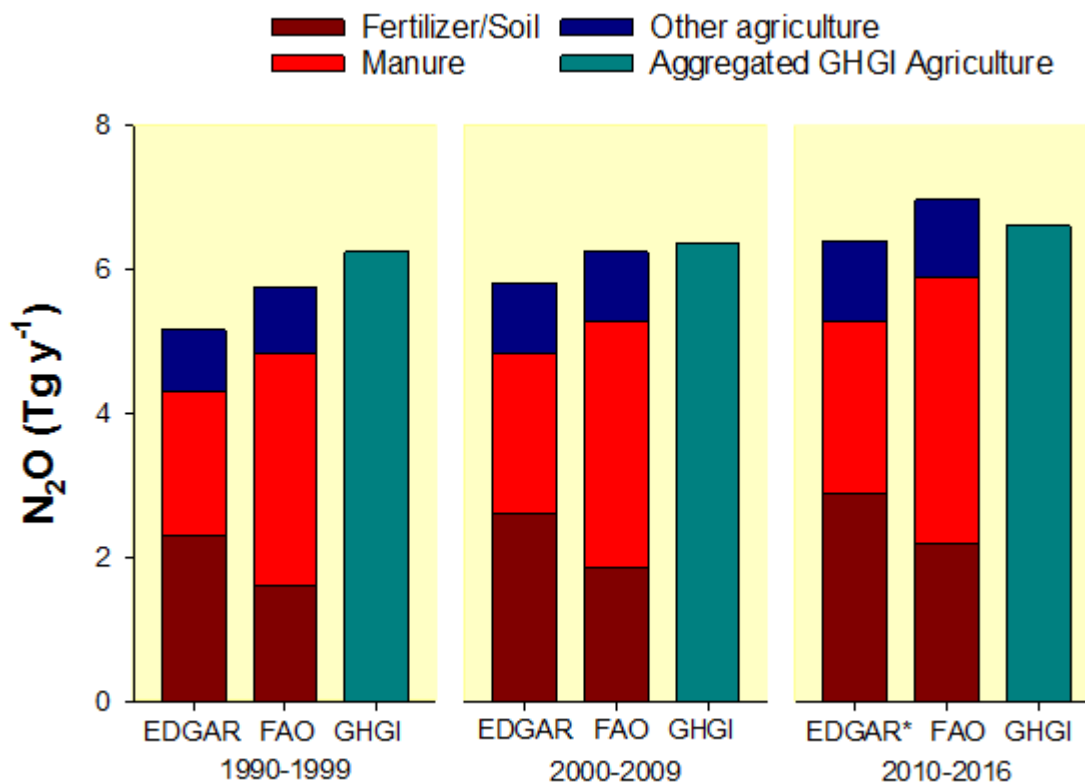


39
 40 **Figure 2.10 Globally averaged atmospheric N₂O mixing ratios since 1984. Data sources: NOAA/ESRL**

1 **Global Monitoring Division (<https://www.esrl.noaa.gov/gmd/hats/combined/N2O.html>).**

2
3 The three independent sources of N₂O emissions estimates from agriculture at global, regional, and national
4 levels are: U.S.E.P.A., EDGAR and FAOSTAT (USEPA 2013; Tubiello et al. 2015; Janssens-Maenhout et
5 al. 2017). EDGAR and FAOSTAT have temporal resolution beyond 2005 and we these databases compare
6 well with national inventory data (Figure 2.10). USEPA has historical estimates through 2005 and
7 projections thereafter. The independent data use IPCC methods, with Tier 1 emission factors and national
8 reporting of activity data. Tier 2 approaches are also available based on top-down and bottom-up
9 approaches. Recent estimates using inversion modelling and process models estimate total annual global
10 N₂O emissions of 16.1-18.7 (bottom-up) and 15.9-17.7 Tg N (top-down), demonstrating relatively close
11 agreement (Thompson et al. 2014). Agriculture is the largest source and has increased with the
12 extensification and intensification. Recent modelling estimates of terrestrial sources show a higher emissions
13 range that is slightly more constrained than what was reported in AR5: approximately 9 (7–11) Tg N₂O-N yr⁻¹
14 (Saikawa et al. 2014; Tian et al. 2016) compared to 6.6 (3.3–9.0) Tg N₂O-N yr⁻¹ (Ciais et al. 2013a).
15 Estimates of marine N₂O emissions are between 2.5 and 4.6 Tg N₂O-N yr⁻¹; (Buitenhuis et al., 2017;
16 Saikawa et al., 2014).

17
18 To conclude, N₂O is continuing to accumulate in the atmosphere at an increasingly higher rate (*very high*
19 *confidence*), driven primarily by increases in manure production and synthetic N fertiliser use from the mid-
20 20th century onwards (*high confidence*). Findings since AR5 have constrained regional and global estimates
21 of annual N₂O emissions and improved our understanding of the spatio-temporal dynamics of N₂O
22 emissions, with soil rewetting and freeze-thaw cycles, which important determinants of total annual emission
23 fluxes in some regions (*medium confidence*).



24
25 **Figure 2.11 Average agricultural N₂O emissions estimates from 1990. Sub-sectorial agricultural emissions**
26 **are based on the Emissions Database for Global Atmospheric Research (EDGAR v4.3.2; Janssens-**
27 **Maenhout et al. 2017a); FAOSTAT (Tubiello et al. 2013); and National GHGI data (Grassi et al. 2018).**
28 **GHGI data are aggregate values for the sector.**

29 * Note that EDGAR data are complete only through 2012; the EDGAR data in the right-hand panel
30 represent the three years 2010-2012 and are presented for comparison.

31 32 2.3.3.2 Land use effects

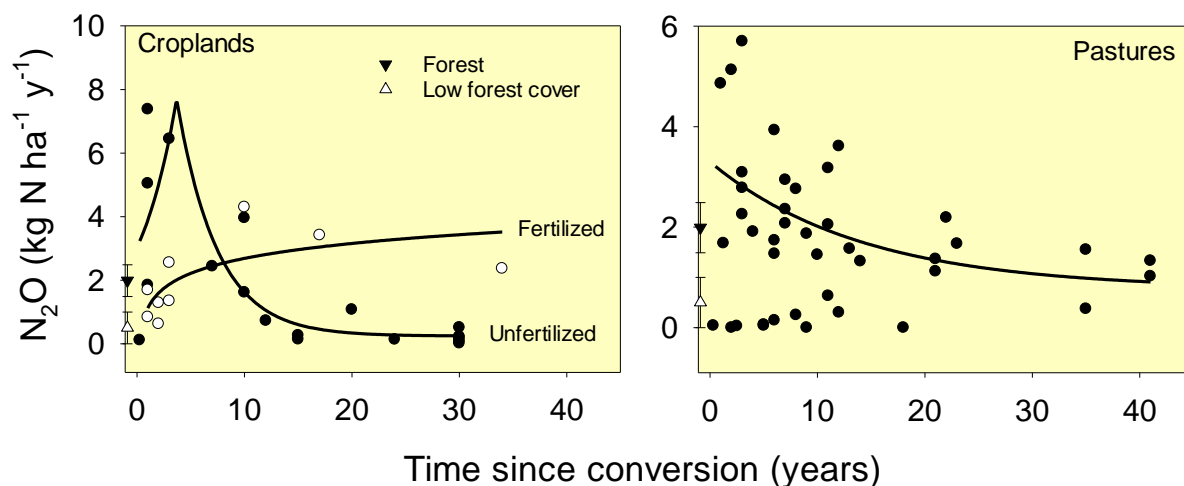
33 Agriculture is responsible for approximately two-thirds of N₂O emissions (*robust evidence, high agreement*)

1 (Janssens-Maenhout et al. 2017). Total emissions from this sector are the sum of direct and indirect
2 emissions. Direct emissions from soils are the result of mineral fertiliser and manure application, manure
3 management, deposition of crop residues, cultivation of organic soils and inorganic N inputs through
4 biological nitrogen fixation. Indirect emissions come from increased warming, enrichment of downstream
5 water bodies from runoff, and downwind N deposition on soils. The main driver of N₂O emissions in
6 croplands is a lack of synchronisation between crop N demand and soil N supply, with approximately 50%
7 of N applied to agricultural land not taken up by the crop (Zhang et al. 2017). Cropland soils emit over 3 Tg
8 N₂O-N yr⁻¹ (*medium evidence, high agreement*) (Janssens-Maenhout et al. 2017; Saikawa et al. 2014).
9 Regional inverse modelling studies show larger tropical emissions than the inventory approaches and they
10 show increases in N₂O emissions from the agricultural sector in South Asia, Central America, and South
11 America (Saikawa et al. 2014; Wells et al. 2018).

12
13 Emissions of N₂O from pasturelands and rangelands have increased by as much as 80% since 1960 due to
14 increased manure production and deposition (*robust evidence, high agreement*) (de Klein et al. 2014; Tian et
15 al. 2018; Chadwick et al. 2018; Dangal et al. 2019; Cardenas et al. 2019). Studies consistently report that
16 pasturelands and rangelands are responsible for around half of the total agricultural N₂O emissions
17 (Davidson 2009; Oenema et al. 2014; Dangal et al. 2019). An analysis by Dangal et al. (2019) shows that
18 while managed pastures make up around one-quarter of the global grazing lands, they contribute 86% of the
19 net global N₂O emissions from grasslands and that more than half of these emissions are related to direct
20 deposition of livestock excreta on soils.

21
22 Many studies calculate N₂O emissions from a linear relationship between nitrogen application rates and N₂O
23 emissions. New studies are increasingly finding nonlinear relationships, which means that N₂O emissions per
24 hectare are lower than the Tier 1 EFs (IPCC 2003) at low nitrogen application rates, and higher at high
25 nitrogen application rates (*robust evidence, high agreement*) (Shcherbak et al. 2014; van Lent et al. 2015;
26 Satria 2017). This not only has implications for how agricultural N₂O emissions are estimated in national and
27 regional inventories, which now often use a linear relationship between nitrogen applied and N₂O emissions,
28 it also means that in regions of the world where low nitrogen application rates dominate, increases in
29 nitrogen fertiliser use would generate relatively small increases in agricultural N₂O emissions. Decreases in
30 application rates in regions where application rates are high and exceed crop demand for parts of the growing
31 season are likely to have very large effects on emissions reductions (*medium evidence, high agreement*).

32
33 Deforestation and other forms of land-use change alter soil N₂O emissions. Typically, N₂O emissions
34 increase following conversion of native forests and grasslands to pastures or croplands (McDaniel et al.
35 2019; van Lent et al. 2015). This increase lasts from a few years to a decade or more, but there is a trend
36 toward decreased N₂O emissions with time following land use change and ultimately lower N₂O emissions
37 than had been occurring under native vegetation, in the absence of fertilisation (**Figure 2.12**) (Meurer et al.
38 2016; van Lent et al. 2015) (*medium evidence, high agreement*). Conversion of native vegetation to fertilised
39 systems typically leads to increased N₂O emissions over time, with the rate of emission often being a
40 function of nitrogen fertilisation rates, but this response can be moderated by soil characteristics and water
41 availability (*medium evidence, high agreement*) (van Lent et al. 2015; Meurer et al. 2016). Restoration of
42 agroecosystems to natural vegetation, over the period of one to two decades does not lead to recovery of N₂O
43 emissions to the levels of the original vegetation (McDaniel et al. 2019). To conclude, findings since AR5
44 increasingly highlight the limits of linear N₂O emission factors, particularly from field to regional scales,
45 with emissions rising nonlinearly at high nitrogen application rates (*high confidence*). Emissions from
46 unfertilised systems often increase and then decline over time with typically lower emissions than was the
47 case under native vegetation (*high confidence*).



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Figure 2.12 Effect of time since conversion on N_2O fluxes in unfertilised (black circles) and fertilised (white circles) tropical croplands (left frame) and in unfertilised tropical pastures (right frame). Average N_2O flux and 95% confidence intervals are given for upland forests (black inverted triangle) and low canopy forests (white inverted triangle), for comparison. The solid lines represent the trends for unfertilised and fertilised cases. Data source: (van Lent et al. 2015).

While soil emissions are the predominant source in agriculture, other sources are important or their importance is only just emerging. Biomass burning is responsible for approximately $0.7 \text{ Tg N}_2\text{O-N yr}^{-1}$ ($0.5\text{--}1.7 \text{ Tg N}_2\text{O-N yr}^{-1}$) or 11% of total gross anthropogenic emissions due to the release of N_2O from the oxidation of organic nitrogen in biomass (van der Werf et al. 2013). This source includes crop residue burning, forest fires, household cook stoves, and prescribed savannah, pasture and cropland burning. Aquaculture is currently not accounted for in most assessments or compilations. While it is currently responsible for less than $0.1 \text{ Tg N}_2\text{O-N yr}^{-1}$, is one of the fastest growing sources of anthropogenic N_2O emissions (Williams and Crutzen 2010; Bouwman et al. 2013) (*limited evidence, high agreement*). Finally, increased nitrogen deposition from terrestrial sources is leading to greater indirect N_2O emissions, particularly since 1980 (*moderate evidence, high agreement*). (Tian et al. 2018, 2016). In marine systems, deposition is estimated to have increased the oceanic N_2O source by $0.2 \text{ Tg N}_2\text{O-N yr}^{-1}$ or 3% of total gross anthropogenic emissions (Suntharalingam et al. 2012).

Box 2.2: Methodologies for estimating national to global scale anthropogenic land carbon fluxes

Bookkeeping/accounting models (Houghton et al. 2012b; Hansis et al. 2015; Houghton and Nassikas 2017) calculate changes in biomass and soils that result from changes in land activity using data on biomass density and rates of growth/decomposition, typically from ground-based inventory data collection (field measurements of carbon in trees and soils). The approach includes only those changes directly caused by major categories of land-use change and management. The models do not explicitly include the indirect effects to changing environmental conditions, although some effects are implicit in biomass, growth rates and decay rates used. Thus, the models may overestimate past fluxes. The bookkeeping models include fluxes from peatland burning based on GFED estimates (Global Fire Emissions Database, (Randerson et al. 2015).)

Dynamic Global Vegetation Models (DGVMs) simulate ecological processes, such as photosynthesis, respiration, allocation, growth, decomposition etc., driven by environmental conditions (climate variability, climate change, CO_2 , nitrogen concentrations). Models vary with respect to the processes included, with many since AR5 now including forest management, fire, N, and other management (Sitch et al. 2005; Le Quéré et al. 2018). Models are forced with increasing atmospheric CO_2 and changing climate, and run with and without “land use change” (land cover and forest harvest) to differentiate the anthropogenic effects from

1 the indirect effects of climate and CO₂ - the “land sink”. Thus, indirect effects are explicitly included. This
2 approach also includes a “lost atmospheric sink capacity”, or the carbon uptake due to environmental effects
3 on forests that does not happen once the forests are removed (Pongratz et al. 2010).
4

5 **Integrated Assessment Models (IAMs)** use story-lines to construct alternative future scenarios of GHG
6 emissions and atmospheric concentrations within a global socio-economic framework, including projections
7 of AFOLU based on assumptions of, for example, crop yields, population growth, bioenergy use (See Cross-
8 Chapter Box 1: Scenarios, Chapter 1). Some models include simplified DGVMs, which may include climate
9 and CO₂ effects, while others use AFOLU emissions from other sources.
10

11 **Earth system models (ESMs)** couple DGVMs, surface hydrology and energy exchange models with a
12 atmosphere, ocean, and sea ice models, enabling exploration of feedbacks between climate change and the
13 carbon cycle (e.g., warming effects increase soil and plant respiration and lead to higher atmospheric CO₂
14 concentrations, which in turn promote plant growth) (Friedlingstein et al. 2014). They sometimes include
15 numerical experiments with and without land-use change to diagnose the anthropogenic AFOLU flux
16 (Lawrence et al. 2016).
17

18 **Satellite data** can be used as a proxy for plant activity (e.g. greenness) and to map land cover, vegetation
19 fires and biomass density. Algorithms, models and independent data are used to calculate fluxes of CO₂ from
20 satellite data, although calculating the net carbon flux is difficult because of the lack of information on the
21 respiratory flux. Some active satellite sensors (LiDAR) are able to measure three-dimensional structure in
22 woody vegetation, which is closely related to biomass density (Zarin et al. 2016a; Baccini et al. 2012;
23 Saatchi et al. 2011). Together with land-cover change data, these estimates of biomass density can be used to
24 provide observational-based estimates of fluxes due to changes in forest area (e.g., (Tyukavina et al. 2015;
25 Harris et al. 2015; Baccini et al. 2012) or degradation (Baccini et al. 2017). Satellite estimates of biomass
26 vary considerably (Mitchard et al. 2013; Saatchi et al. 2015; Avitabile et al. 2016); data are available only for
27 recent decades; methods generally assume that all losses of carbon are immediately released to the
28 atmosphere; and changes in soil carbon are generally ignored. The approach implicitly includes indirect and
29 natural disturbance effects as well as direct anthropogenic effects.
30

31 **Atmospheric Inversions** use observations of atmospheric concentrations with a model of atmospheric
32 transport, based on data for wind speed and direction, to calculate implied emissions (Gatti et al. 2014; Liu et
33 al. 2017a; van der Laan-Luijkx et al. 2017). Since AR5 there has been an increase in availability of
34 concentration data from flux tower networks and satellites, enabling better global coverage at finer spatial
35 scales and some national estimates (e.g. in the UK inverse techniques are used together with national GHG
36 inventories). A combination of concentrations of different gases and isotopes enables the separation of fossil,
37 ocean and land fluxes. However, inversions give only the net flux of CO₂ from land; they cannot separate
38 natural and anthropogenic fluxes.
39

40 **Micrometeorological flux measurements:** Data on CO₂ concentrations and air movements recorded on
41 instrumented towers enable calculation of CO₂ flux at the ecosystem scale. Global and regional Flux
42 Networks (FluxNet (Global), AsiaFlux, Ameriflux (North America), ICOS (EU), NEON (USA), and others)
43 contribute to a global flux data base, which is used to verify the results of modelling, inventory and remote
44 sensing studies.
45

46 **FAOSTAT:** The United Nations Food and Agricultural Organization has produced country level estimates
47 of greenhouse gas emissions (Tubiello et al. 2013) from agriculture (1961–2016) and land use (1990–2016)
48 using a globally consistent methodological approach based largely on IPCC Tier 1 methods of the 2006
49 IPCC Guidelines (FAO 2015b). FAO emissions estimates were used as one of the three database inputs into
50 the AR5 WGIII AFOLU chapter. Non-CO₂ emissions from agriculture are estimated directly from national
51 statistics of activity data reported by countries to FAO. CO₂ emissions from land use and land-use change are
52 computed mostly at Tier 1, albeit at fine geospatial scales to capture effects from peatland degradation and
53 biomass fires (Rossi et al. 2016). Emissions from forest land and deforestation are based on the IPCC carbon
54 stock change method, thus constituting a Tier 3 estimate relying on country statistics of carbon stocks and
55 forest area collected through the FAO FRA. The carbon flux is estimated assuming instantaneous emissions
56 in the year of forest area loss, and changes in carbon stocks within extant forests, but does not distinguish

1 “managed” and “unmanaged” forest areas, albeit it treats separately emissions from primary, secondary and
 2 planted forest (Federici et al. 2015).
 3

4 **Country Reporting of GHG Inventories (GHGIs):** All Parties to the UNFCCC are required to report
 5 national GHG Inventories (GHGIs) of anthropogenic emissions and removals. Reporting requirements are
 6 differentiated between developed and developing countries. Because of the difficulty of separating direct
 7 anthropogenic fluxes from indirect or natural fluxes, the (IPCC 2003) adopted the “managed land” concept
 8 as a proxy to facilitate GHGI reporting. All GHG fluxes on “managed land” are defined as anthropogenic,
 9 with each country applying their own definition of “managed land” (i.e. “where human interventions and
 10 practices have been applied to perform production, ecological or social functions” (IPCC 2006)). Fluxes may
 11 be determined on the basis of changes in carbon stocks (e.g., from forest inventories) or by activity data (e.g.
 12 area of land cover change management activity multiplied by emission factors or with modelled fluxes).
 13 Depending on the specific methods used, GHGIs include all direct anthropogenic effects and may include the
 14 indirect anthropogenic effects of environmental change (generally sinks) and natural effects (see Section
 15 2.3.1.2). GHG fluxes from “unmanaged land” are not reported in GHGIs because they are assumed to be
 16 non-anthropogenic. The reported estimates may then be filtered through agreed “accounting rules” - i.e. what
 17 countries actually count towards their mitigation targets (Cowie AL et al. 2007; Lee, D. and Sanz 2017). The
 18 accounting aims to better quantify the additional mitigation actions by, for example, factoring out the impact
 19 of natural disturbances and forest age-related dynamics (Canadell et al. 2007; Grassi et al. 2018).
 20

25 **Box 2.3: CO₂ fertilisation and enhanced terrestrial uptake of carbon**

26
 27 All Dynamic Global Vegetation Models (DGVMs) and Earth System Models (ESMs) represent the CO₂
 28 fertilisation effect ((Le Quéré et al. 2017; Hoffman et al. 2014). There is *high confidence* that elevated CO₂
 29 results in increased short-term CO₂ uptake per unit leaf area (Swann et al. 2016; Field et al. 1995; Donohue
 30 et al. 2013);. However, whether this increased CO₂ uptake at the leaf level translates into increased growth
 31 for the whole plant differs among plant species and environments because growth is constrained by whole-
 32 plant resource allocation, nutrient limitation (e.g., nitrogen (N), phosphorus (P), potassium (K) and soil water
 33 and light limitations (Körner 2006; Peñuelas et al. 2017; Friend et al. 2014a). Interactions between plants and
 34 soil microbes, further modulate the degree of nutrient limitation on CO₂ fertilisation (Terrer et al. 2017).
 35

36 At the ecosystems level, enhanced CO₂ uptake at decadal or longer time scales depends on changes in plant
 37 community composition and ecosystem respiration, as well disturbance and natural plant mortality (De
 38 Kauwe et al., 2016; Farrior et al., 2015; Keenan et al., 2017; Sulman et al, 2019). The results of FACE
 39 experiments (free-air carbon dioxide enrichment) over two decades are highly variable because of these
 40 factors (Norby et al. 2010; Körner 2015; Feng et al. 2015; Paschalis et al. 2017; Terrer et al. 2017; Du et al.
 41 2019b). Under higher atmospheric CO₂ concentrations, the ratio of CO₂ uptake to water loss (water use
 42 efficiency, WUE), increases and enhances drought tolerance of plants (*high confidence*) (Berry et al., 2010;
 43 Ainsworth & Rogers, 2007).
 44

45 Long-term CO₂ and water vapour flux measurements show that WUE in temperate and boreal forests of the
 46 Northern Hemisphere has increased more than predicted by photosynthetic theory and models over the past
 47 two decades (*high confidence*) (Keenan et al. 2013; Laguë and Swann 2016b). New theories have emerged
 48 on how CO₂ uptake by trees is related to water loss and to the risk of damaging xylem (water conducting
 49 tissues) in the trunk and branches (Wolf et al. 2016a; Anderegg et al. 2018a). Tree ring studies of stable
 50 carbon and oxygen isotopes also detected increased WUE in recent decades (Battipaglia et al. 2013; Silva
 51 and Anand 2013; van der Sleen et al. 2014). Yet, tree ring studies often fail to show acceleration of tree
 52 growth rates in support of CO₂ fertilisation, even when they show increased WUE (van der Sleen et al.
 53 2014). The International Tree Ring Data Bank (ITRDB) indicated that only about 20% of the sites in the
 54 database showed increasing trends in tree growth that cannot be explained by climate variability, nitrogen
 55 deposition, elevation, or latitude. Thus there is *limited evidence (low agreement)* among observations of
 56 enhanced tree growth due to CO₂ fertilisation of forests during the 20th century (Gedalof and Berg 2010).

1
2 In grasslands, although it is possible for CO₂ fertilisation to alleviate the impacts of drought and heat stress
3 on net carbon uptake (Roy et al. 2016b), there is *low confidence* about its projected magnitude. Because of
4 its effect on water use efficiency, CO₂ fertilisation is expected to be pronounced in semi-arid habitats; and
5 because of different metabolic pathways, C₃ plants are expected to be more sensitive to elevated CO₂
6 concentrations than C₄ grasses (Donohue et al. 2013; Morgan et al. 2011; Derner et al. 2003). Neither of
7 these expectations was observed over a 12-year study of elevated CO₂ in a grassland system: enhanced
8 growth was not observed during dry summers, and growth of C₄ grasses was unexpectedly stimulated, while
9 growth of C₃ grasses was not (Reich et al. 2014, 2018).

10
11 There is *medium confidence* that CO₂ fertilisation effects have increased water use efficiency in crops and
12 thus reduced agricultural water use per unit of crop produced (Deryng et al. 2016b; Nazemi and Wheeler
13 2015; Elliott et al. 2014) . This effect could lead to near-term continued greening of agricultural areas.
14 However, current assessments of these effects are based on limited observations, mostly from the temperate
15 zone (Deryng et al. 2016a).

16
17 One line of evidence for CO₂ fertilisation is the increasing land sink (“the residual land sink” in AR5) over
18 the last 50 years as the atmospheric CO₂ concentration has increased (Los 2013; Sitch et al. 2015b; Campbell
19 et al. 2017; Keenan and Riley 2018). A combined analysis of atmospheric inverse analyses, ecosystem
20 models, and forest inventory data concluded that 60% of the recent terrestrial carbon sink can be directly
21 attributed to increasing atmospheric CO₂ (Schimel et al. 2015). A global analysis using a “reconstructed
22 vegetation index” (RVI) for the period 1901–2006 from MODIS satellite-derived NDVI (Normalised
23 Vegetation Difference Index) showed that CO₂ fertilisation contributed at least 40% of the observed increase
24 in the land carbon sink (Los 2013). Without CO₂ fertilisation ESMs are unable to simulate the increasing
25 land sink and the observed atmospheric CO₂ concentration growth rate since the middle of the 20th century
26 (Shevliakova et al. 2013). There are other mechanisms that could explain enhanced land C uptake such as
27 increased regional forest and shrub cover (see Cross-Chapter Box 2: Implications of large-scale conversion
28 from non-forest to forest land, Chapter 1);(Chen et al. 2019), and, at higher latitudes, increasing temperatures
29 and longer growing seasons (Zhu et al. 2016).

30
31 In summary, there is *low confidence* about the magnitude of the CO₂ effect and other factors that may
32 explain at least a portion of the land sink (e.g., nitrogen deposition, increased growing season, reduced
33 burning, erosion and re-deposition or organic sediments, and aerosol-induced cooling). Increases in
34 atmospheric CO₂ result in increased water use efficiency and increase leaf-level photosynthesis (*high*
35 *confidence*). The extent to which CO₂ fertilisation results in plant- or ecosystem-level carbon accumulation is
36 highly variable and affected by other environmental constraints (*high confidence*). Even in ecosystems where
37 CO₂ fertilisation has been detected in recent decades, those effects are found to weaken as a result of
38 physiological acclimation, soil nutrient limitation, and other constraints on growth (Friend et al., 2014;
39 Körner, 2006; Peñuelas et al., 2017).

40 41 42 **2.4 Emissions and impacts of short-lived climate forcers (SLCF) from land**

43
44 While the rising atmospheric concentration of GHGs is the largest driver of anthropogenic changes in
45 climate, the levels of short-lived climate forcers (SLCF) can significantly modulate regional climate by
46 altering radiation exchanges and hydrological cycle and impact ecosystems (Boucher et al. 2013; Rogelj et
47 al. 2014; Kok et al. 2018) (*high confidence*). This section assesses the current state of knowledge with
48 respect to past and future emissions of the three major SLCFs and their precursors: mineral dust,
49 carbonaceous aerosols (Black Carbon and Organic Carbon), and Biogenic volatile organic compounds
50 (BVOCs). The chapter also reports on implications of changes in their emissions for climate. Aerosols
51 particles with diameters between about 0.010 µm to about 20 µm are recognised as SLCFs, a term that refers
52 to their short atmospheric lifetime (a few days). BVOCs are important precursors of ozone and organic
53 carbon (OC), both important climate forcing agents with short atmospheric lifetimes.

54
55 While the AR5 did not assess land aerosols emissions in depth, their findings stated that although progress in
56 quantifying regional emissions of anthropogenic and natural land aerosols has been made, considerable

1 uncertainty still remains about their historical trends, their inter-annual and decadal variability and about any
2 changes in the future (Calvo et al. 2013; Klimont et al. 2017). Some new and improved understanding of
3 processes controlling emissions and atmospheric processing has been developed since AR5, for example, a
4 better understanding of the climatic role of Black Carbon (BC) as well as the understanding of the role of
5 BVOCs in formation of secondary organic aerosols (SOA).

6
7 Depending on the chemical composition and size, aerosols can absorb or scatter sunlight and thus directly
8 affect the amount of absorbed and scattered radiation (Fuzzi et al. 2015a; Nousiainen 2011; de Sá et al. 2019)
9 Aerosols affect clouds formation and development, and thus can also influence precipitation patterns and
10 amounts (Sun et al. 2015). In addition, deposition of aerosols—especially black carbon—on snow and ice
11 surfaces can reduce albedo and increase warming as a self-reinforcing feedback. Aerosols deposition also
12 change biogeochemical cycling in critical terrestrial ecosystems with deposition of nutrients such as nitrogen
13 and phosphorus (Andreae et al. 2002). Primary land aerosols are emitted directly into the atmosphere due to
14 natural or anthropogenic processes and include mineral aerosols (or dust), volcanic dust, soot from
15 combustion, organic aerosols from industry, vehicles or biomass burning, bioaerosols from forested regions,
16 and others. Secondary organic aerosols (SOA) are particulates that are formed in the atmosphere by gas-to-
17 particles conversion processes from gaseous precursors, such as BVOCs, and account for a large fraction of
18 fine mode (particles less than 2.5µm) aerosol mass (Hodzic et al. 2016; Manish et al. 2017). Land use change
19 can affect the climate through changed emissions of SLCFs such as aerosols, ozone precursors and methane.

20
21 Aerosols from air pollution will decline in the coming years as a means for improving urban and regional air,
22 but their removal will lead to additional warming (Boucher et al. 2013), with important regional variability,
23 and partially offsetting projected mitigation effects for two to three decades in 1.5°C consistent pathways
24 (IPCC 2018) (*high confidence*). It is important to emphasise that changes in emissions can either be due to
25 external forcing or through a feedback in the climate system (Box 2.1:). For instance, enhanced dust
26 emissions due to reduced vegetation could be a forcing if overgrazing is the cause of larger dust emission, or
27 a feedback if dryer climate is the cause. This distinction is important in terms of mitigation measures to be
28 implemented.

29 30 **2.4.1 Mineral dust**

31
32 One of the most abundant atmospheric aerosols emitted into the atmosphere is mineral dust, a “natural”
33 aerosol that is produced by wind strong enough to initiate the emissions process of sandblasting. Mineral
34 dust is preferentially emitted from dry and unvegetated soils in topographic depressions where deep layer of
35 alluvium have been accumulated (Prospero et al. 2002). Dust is also emitted from disturbed soils by human
36 activities with a 25% contribution to global emission, based on satellite-based estimate (Ginoux et al. 2012).
37 Dust is then transported over long distances across continents and oceans. Dust cycle, which consists of
38 mineral dust emission, transport, deposition and stabilisation, have multiple interactions with many climate
39 processes and biogeochemical cycles.

40 41 **2.4.1.1 Mineral dust as a short-lived climate forcer from land**

42 Depending on the dust mineralogy, mixing state, and size, dust particles can absorb or scatter shortwave and
43 long-wave radiation. Dust particles serve as cloud condensation nuclei and ice nuclei. They can influence the
44 microphysical properties of clouds, their lifetime and precipitation rate (Kok et al. 2018). New and improved
45 understanding of processes controlling emissions and transport of dust, its regional patterns and variability as
46 well as its chemical composition has been developed since AR5.

47
48 While satellites remain the primary source of information to locate dust sources and atmospheric burden, in-
49 situ data remains critical to constrain optical and mineralogical properties of the dust (Di Biagio et al. 2017;
50 Rocha-Lima et al. 2018). Dust particles are composed of minerals, including iron oxides which strongly
51 absorb shortwave radiation and provide nutrient for marine ecosystems. Other mineral such as feldspar is an
52 efficient ice nuclei (Harrison et al. 2016). Dust mineralogy depending on the native soils, global databases
53 were developed to characterise mineralogical composition of soils for use in the weather and climate models
54 (Journet et al. 2014; Perlwitz et al. 2015). New field campaigns as well as new analysis from prior campaign
55 have produced insights into role of dust in western Africa in climate system, for example, for dust
56 (Veselovskii et al. 2016), long-ranged transport of dust across the Atlantic (Groß et al. 2015), and the

1 characterisation of aerosol particles and their ability to act as ice and cloud condensation nuclei (Price et al.
2 2018). Size distribution at emission is another key parameter controlling dust interactions with radiation.
3 Most models use now the parametrisation of Kok (2011) based on the theory of brittle material. It was shown
4 that most models underestimate the size of global dust cycle (Kok 2011) has been underestimated.
5 Characterisation of spatial and temporal distribution of dust emissions is essential for weather prediction and
6 climate projections (*high confidence*). Although there is a growing confidence in characterising the
7 seasonality and peak of dust emissions (i.e., spring-summer, (Wang et al. 2015)) and how the meteorological
8 and soil conditions control dust sources, an understanding of long-term future dust dynamics, inter-annual
9 dust variability and how they will affect future climate still requires substantial work. Dust is also important
10 at high latitude, where it has impacts on snow covered surface albedo and weather (Bullard et al. 2016).

11 **2.4.1.2 Effects of past climate change on dust emissions, and feedbacks**

12 Limited number of model-based studies found that dust emissions increased significantly since late 19th
13 century: by 25% from preindustrial to present day (e.g., from 729 Tg yr⁻¹ to 912 Tg yr⁻¹) with ~50% increase
14 driven by climate change and ~40% by land use cover change such as conversion of natural land to
15 agriculture (Stanelle et al. 2014) (*low confidence*). These changes resulted in a clear sky radiative forcing at
16 the top of the atmosphere of -0.14 W m⁻² (Stanelle et al. 2014). The authors found that, in North Africa most
17 dust is of natural origin with a recent 15% increase in dust emissions attributed to climate change; in North
18 America two thirds of dust emissions take place on agricultural lands and both climate change and land use
19 change jointly drive the increase; between pre-industrial and present-day the overall effect of changes in dust
20 is -0.14 W m⁻² cooling of clear sky net radiative forcing on top of the atmosphere, with -0.05 W m⁻² from land
21 use and -0.083 W m⁻² from changes in climate.

22
23
24 The comparison of observations for vertically integrated mass of atmospheric dust mass per unit area (i.e.
25 Dust Mass Path or DMP) obtained from the remotely sensed data and the DMP from CMIP5 models reveal
26 that model-simulate range of DMP was much lower than the estimates from (Evan et al. 2014). ESM
27 typically do not reproduce inter-annual and longer time scales variability seen in observations (Evan et al.
28 2016). Analyses of the CMIP5 models (Evan 2018; Evan et al. 2014)) reveal that all climate models
29 systematically under-estimate dust emissions, amount of dust in the atmosphere and its inter-annual
30 variability (*medium confidence*).

31
32 One commonly suggested reason for the lack of dust variability in climate models is the models' inability to
33 simulate the effects of land surface changes on dust emission (Stanelle et al. 2014). Models which account
34 for changes in land surface show more agreement with the satellite observations both in terms of Aerosol
35 Optical Depth and DMP (Kok et al. 2014). New prognostic dust emissions models now able to account for
36 both changes in surface winds and vegetation characteristics (e.g., leaf area index and stem area index) and
37 soil water, ice, and snow cover (Evans et al. 2016). As a result, new modelling studies (e.g. Evans et al.
38 2016) indicate that in regions where soil and vegetation respond strongly to ENSO events, such as in
39 Australia, inclusion of dynamic vegetation characteristics into dust emission parameterisations improves
40 comparisons between the modelled and observed relationship long-term climate variability (e.g., ENSO) and
41 dust levels (Evans et al. 2016). Thus, there has been progress in incorporating effects of vegetation, soil
42 moisture, surface wind and vegetation on dust emission source functions but the number of studies
43 demonstrating such improvement remains small (*limited evidence, medium agreement*).

44 **2.4.1.3 Future changes of dust emissions**

45
46
47 There is no agreement about direction of future changes in dust emissions. Atmospheric dust loading is
48 projected to increase over the southern edge of the Sahara in association with surface wind and precipitation
49 changes (Pu and Ginoux, 2018), while Evan et al. (2016) project a decline in African dust emissions. Dust
50 Optical Depth (DOD) is also projected to increase over the central Arabian Peninsula in all seasons and to
51 decrease over northern China from MAM to SON (Pu and Ginoux 2018). Climate models project rising
52 drought risks over the southwestern and central US. in the twenty-first century. The projected drier regions
53 largely overlay the major dust sources in the US. However, whether dust activity in the US will increase in
54 the future is not clear, due to the large uncertainty in dust modelling (Pu and Ginoux 2017). Future trends of
55 dust emissions will depend on changes in precipitation patterns and atmospheric circulation (*limited
56 evidence, high agreement*). However, implication of changes in human activities, including mitigation (e.g.

1 bioenergy production) and adaption (e.g. irrigation) are not characterised in the current literature.

2 3 4 **2.4.2 Carbonaceous Aerosols**

5
6 Carbonaceous aerosols are one of the most abundant components of aerosol particles in continental areas of
7 the atmosphere and a key land-atmosphere component (Contini et al. 2018). They can make up to 60-80% of
8 PM_{2.5} (Particulate matter with size less than 2.5 µm) in urban and remote atmosphere (Tsigaridis et al.
9 2014a; Kulmala et al. 2011). It comprises an organic fraction (Organic Carbon - OC) and a refractory light
10 absorbing component, generally referred as Elemental Carbon (EC), from which Black Carbon (BC) is the
11 optically active absorption component of EC (Gilardoni et al. 2011; Bond et al. 2013).

12 13 **2.4.2.1 Carbonaceous aerosol precursors of short-lived climate forcers from land**

14 OC is a major component of aerosol mass concentration, and it originates from different anthropogenic
15 (combustion processes) and natural (from natural biogenic emissions) sources (Robinson et al. 2007). A
16 large fraction of OC in the atmosphere has a secondary origin, as it can be formed in the atmosphere through
17 condensation to the aerosol phase of low vapour pressure gaseous compounds emitted as primary pollutants
18 or formed in the atmosphere. This component is called Secondary Organic Aerosol (SOA) (Hodzic et al.
19 2016). A third component of the optically active aerosols is the so-called brown carbon (BrC), an organic
20 material that shows enhanced solar radiation absorption at short wavelengths (Wang et al. 2016b; Laskin et
21 al. 2015; Liu et al. 2016a; Bond et al. 2013; Saturno et al. 2018).

22
23 OC and EC have distinctly different optical properties, with OC being important for the scattering properties
24 of aerosols and EC is central for the absorption component (Rizzo et al. 2013; Tsigaridis et al. 2014a; Fuzzi
25 et al. 2015a). While organic carbon is reflective and scatter solar radiation, it has a cooling effect on climate.
26 On the other side, BC and BrC absorbs solar radiation and they have a warming effect in the climate system.
27 (Bond et al. 2013).

28
29 Organic carbon is also characterised by a high solubility with a high fraction of water-soluble organic
30 compounds (WSOC) and it is one of the main drivers of the oxidative potential of atmospheric particles. This
31 makes particles loaded with oxidised OC an efficient CCN in most of the conditions (Pöhlker et al. 2016;
32 Thalman et al. 2017; Schmale et al. 2018).

33
34 Biomass burning is a major global source of carbonaceous aerosols (Bowman et al. 2011b; Harrison et al.
35 2010; Reddington et al. 2016; Artaxo et al. 2013). As knowledge of past fire dynamics improved through
36 new satellite observations, new fire proxies' datasets (Marlon et al. 2013; van Marle et al. 2017), and
37 process-based models (Hantson et al. 2016), a new historic biomass burning emissions dataset starting in
38 1750 has been developed (Van Marle et al. 2017a) (see Cross-Chapter Box 3: Fire and Climate Change, in
39 this chapter). Revised versions of OC biomass burning emissions (Van Marle et al. 2017a) show in general
40 reduced trends compared to the emissions derived by (Lamarque et al. 2010) for CMIP5. CMIP6 global
41 emissions pathways (Gidden et al. 2018; Hoesly et al. 2018) estimate global BC emissions in 2015 at 9.8 Mt
42 BC yr⁻¹, while global OC emissions are 35 Mt OC yr⁻¹.

43
44 Land use change is critically important for carbonaceous aerosols, since biomass burning emissions consist
45 mostly of organic aerosol, and the undisturbed forest is also a large source of organic aerosols (Artaxo et al.
46 2013). Additionally, urban aerosols are also mostly carbonaceous, because of the source composition (traffic,
47 combustion, industry, etc.) (Fuzzi et al. 2015b). Burning of fossil fuel, biomass burning emissions and SOA
48 from natural BVOC emissions are the main global sources of carbonaceous aerosols. Any change in each of
49 these components influence directly the radiative forcing (Contini et al. 2018; Boucher et al. 2013; Bond et
50 al. 2013).

51
52 One important component of carbonaceous aerosols is the primary biological aerosol particles (PBAP), also
53 called bioaerosols, that correspond to a significant fraction of aerosols in forested areas (Fröhlich-Nowoisky
54 et al. 2016; Pöschl and Shiraiwa 2015). They are emitted directly by the vegetation as part of the biological
55 processes (Huffman et al. 2012). Airborne bacteria, fungal spores, pollen, archaea, algae, and other
56 bioparticles are essential for the reproduction and spread of organisms across various terrestrial ecosystems.

1 They can serve as nuclei for cloud droplets, ice crystals, and precipitation, thus influencing the hydrological
2 cycle and climate (Whitehead et al. 2016; Scott et al. 2015; Pöschl et al. 2010).

3 4 **2.4.2.2 Effects of past climate change on carbonaceous aerosols emissions, and feedbacks**

5 Annual global emission estimates of BC range from 7.2-7.5 Tg yr⁻¹ using bottom-up inventories (Bond et al.
6 2013; Klimont et al. 2017) up to 17.8 ± 5.6 Tg yr⁻¹ using a fully coupled climate-aerosol-urban model
7 constrained by aerosol measurements (Cohen and Wang 2014), with considerably higher BC emissions for
8 Eastern Europe, Southern East Asia, and Southeast Asia mostly due to higher anthropogenic BC emissions
9 estimates. A significant source of BC, the net trend in global burned area from 2000 to 2012 was a modest
10 decrease of 4.3 Mha yr⁻¹ (-1.2% yr⁻¹).

11
12 Carbonaceous aerosols are important in urban areas as well as pristine continental regions, since they can be
13 responsible for 50-85% of PM_{2.5} (Contini et al. 2018; Klimont et al. 2017). In boreal and tropical forests,
14 carbonaceous aerosols originate from BVOC oxidation (Section 2.4.3). The largest global source of BC
15 aerosols is open burning of forests, savannah and agricultural lands with emissions of about 2,700 Gg yr⁻¹ in
16 the year 2000 (Bond et al. 2013).

17
18 ESMS most likely underestimate globally averaged EC emissions (Bond et al. 2013; Cohen and Wang 2014),
19 although recent emission inventories have included an upwards adjustment in these numbers (Hoesly et al.
20 2018). Vertical EC profiles have also been shown to be poorly constrained (Samset et al. 2014a), with a
21 general tendency of too much EC at high altitudes. Models differ strongly in the magnitude and importance
22 of the coating-enhancement of ambient EC absorption (Boucher et al. 2016) (Gustafsson and Ramanathan
23 2016), in their estimated lifetime of these particles, as well as in dry and wet removal efficiency (Mahmood
24 et al. 2016) (*limited evidence, medium agreement*).

25
26 The equilibrium in emissions and concentrations between the scattering properties of organic aerosol versus
27 the absorption component of BC is a key ingredient in the future climatic projections of aerosol effects
28 (*limited evidence, high agreement*). The uncertainties in net climate forcing from BC rich sources are
29 substantial, largely due to lack of knowledge about cloud interactions with both black carbon and co-emitted
30 organic carbon. A strong positive forcing of about 1.1 W m⁻² was calculated by (Bond et al. 2013), but this
31 forcing is balanced by a negative forcing of -1.45 W m⁻², and shows clearly a need to work on the co-
32 emission issue for carbonaceous aerosols. The forcing will also depend on the aerosol-cloud interactions,
33 where carbonaceous aerosol can be coated and change their CCN capability. It is difficult to estimate the
34 changes in any of these components in a future climate, but this will influence strongly the radiative forcing
35 (Contini et al. 2018; Boucher et al. 2013; Bond et al. 2013) (*high confidence*).

36
37 De Coninck et al. (2018) reported studies estimating a lower global temperature effect from BC mitigation
38 (e.g., Samset et al. 2014b; Boucher et al. 2016), although commonly used models do not capture properly
39 observed effects of BC and co-emissions on climate (e.g., (Bond et al. 2013). Regionally, the warming
40 effects can be substantially larger, for example, in the Arctic (Sand et al. 2015) and high mountain regions
41 near industrialised areas or areas with heavy biomass burning impacts (Ming et al. 2013) (*high confidence*).

42 43 **2.4.2.3 Future changes of carbonaceous aerosol emissions**

44 Due to the short atmospheric lifetime of carbonaceous aerosols in the atmosphere, of the order of a few days,
45 most studies dealing with the future concentration levels have a regional character (Cholakian et al. 2018;
46 Fiore et al. 2012). The studies agree that the uncertainties in changes in emissions of aerosols and their
47 precursors are generally higher than those connected to climate change itself. Confidence in future changes
48 in carbonaceous aerosol concentration projections is limited by the reliability of natural and anthropogenic
49 emissions (including wildfires, largely caused by human activity) of primary aerosol as well as that of the
50 precursors. The Aerosol Chemistry Model Intercomparison Project (AerChemMIP) is endorsed by the
51 Coupled-Model Intercomparison Project 6 (CMIP6) and is designed to quantify the climate impacts of
52 aerosols and chemically- reactive gases (Lamarque et al. 2013). These simulations calculated future
53 responses to SLCF emissions for the RCP scenarios in terms of concentration changes and radiative forcing.
54 Carbonaceous aerosol emissions are expected to increase in the near future due to possible increases in open
55 biomass burning (forest, savannah, and agricultural fires) emissions, and increase in SOA from oxidation of
56 BVOCs (Tsigaridis et al. 2014b; Van Marle et al. 2017b; Giglio et al. 2013) (*medium confidence*).

1
2 More robust knowledge has been produced since the conclusions reported in AR5 (Boucher et al. 2013) and
3 all lines of evidence now agree on a small effect on carbonaceous aerosol global burden due to climate
4 change (*medium confidence*). The regional effects, however, are predicted to be much higher (Westervelt et
5 al. 2015). With respect to possible changes in the chemical composition of PM as a result of future climate
6 change only a few sparse data are available in the literature and the results are, as yet, inconclusive. The co-
7 benefits of reducing aerosol emissions due to air quality issues will play an important role in future
8 carbonaceous aerosol emissions (Gonçalves et al. 2018; Shindell et al. 2017) (*high confidence*).
9

10 **2.4.3 Biogenic Volatile Organic Compounds (BVOCs)**

11
12 Biogenic volatile organic compounds (BVOCs) are emitted in large amounts by forests (Guenther et al.
13 2012). They include isoprene, terpenes, alkanes, alkenes, alcohols, esters, carbonyls and acids (Peñuelas and
14 Staudt 2010; Guenther et al. 1995, 2012). Their emissions represent a carbon loss to the ecosystem, which
15 can represent up to 10% of the carbon fixed by photosynthesis under stressful conditions (Bracho-Nunez et
16 al. 2011). The global average emission for vegetated surfaces is $0.7\text{ g C m}^{-2}\text{ yr}^{-1}$ but can exceed 100 g C m^{-2}
17 yr^{-1} in some tropical ecosystems (Peñuelas and Llusà 2003).
18

19 **2.4.3.1 BVOC precursors of short-lived climate forcers from Land**

20 BVOCs are rapidly oxidised in the atmosphere to form less volatile compounds that can condense and form
21 secondary organic aerosol (SOA). In boreal and tropical forests, carbonaceous aerosols originate from
22 BVOC oxidation, of which isoprene and terpenes are the most important precursors (Claeys et al. 2004; Hu
23 et al. 2015; De Sá et al. 2017; de Sá et al. 2018; Liu et al. 2016b, see following sub-section). BVOCs are the
24 most important precursors of SOA. This transformation process of BVOCs affects the aerosol size
25 distribution both by contributing to new particle formation and to the growth of larger pre-existing particles.
26 SOA affect the scattering of radiation by the particles themselves (direct aerosol effect), but also change the
27 amount of cloud condensation nuclei (CCN) and the lifetime and optical properties of clouds (indirect
28 aerosol effect).
29

30 High amounts of SOA are observed over forest areas, in particular in boreal and tropical regions where they
31 have been found to mostly originate from BVOC emissions (Manish et al. 2017). In particular, isoprene
32 epoxydiol-derived SOA (IEPOX-SOA) is being identified in recent studies in North America and
33 Amazonian forest as a major component in the oxidation of isoprene (Allan et al. 2014; Schulz et al. 2018;
34 De Sá et al. 2017). In tropical regions BVOC can be convected up to the upper atmosphere, where their
35 volatility is reduced and where they become SOA. In some cases those particles are even transported back to
36 the lower atmosphere (Schulz et al. 2018; Wang et al. 2016a; Andreae et al. 2018). In the upper troposphere
37 in the Amazon, SOA are important CCN and are responsible for the vigorous hydrological cycle (Pöhlker et
38 al. 2018). This strong link between BVOC emissions by plants and hydrological cycle has been discussed in
39 a number of studies (Fuentes et al. 2000; Schmale et al. 2018; Pöhlker et al. 2018, 2016).
40

41 Changing BVOC emissions also affect the oxidant concentrations in the atmosphere. Their impact on the
42 concentration of ozone depends on the NO_x concentrations. In polluted regions, high BVOC emissions lead
43 to increased production of ozone, followed by the formation of more OH and a reduction in the methane
44 lifetime. In more pristine regions (NO_x-limited), increasing BVOC emissions instead lead to decreasing OH
45 and ozone concentrations, resulting in a longer methane lifetime. The net effect of BVOCs then can change
46 over time if NO_x emissions are changing.
47

48 BVOCs' possible climate effects have received little attention because it was thought that their short lifetime
49 would preclude them from having any significant direct influence on climate (Unger 2014a; Sporre et al.
50 2018). Higher temperatures and increased CO₂ concentrations are (separately) expected to increase the
51 emissions of BVOCs (Jardine et al. 2011, 2015; Fuentes et al. 2016). This has been proposed to initiate
52 negative climate feedback mechanisms through increased formation of SOA (Arneth et al. 2010; Kulmala
53 2004; Unger et al. 2017). More SOA can make the clouds more reflective, which can provide a cooling.
54 Furthermore, the increase in SOA formation has also been proposed to lead to increased aerosol scattering,
55 resulting in an increase in diffuse radiation. This could boost gross primary production (GPP) and further
56 increase BVOC emissions (Kulmala et al. 2014; Cirino et al. 2014; Sena et al. 2016; Schafer et al. 2002;

Ometto et al. 2005; Oliveira et al. 2007). These important feedbacks are starting to emerge (Sporre et al. 2018; Kulmala 2004; Arneth et al. 2017b). However, there is evidence that this influence might be significant at different spatial scales, from local to global, through aerosol formation and through direct and indirect greenhouse effects (*limited evidence, medium agreement*). Most tropical forest BVOC are primarily emitted from foliage of trees but soil microbes can also be a major source of some compounds including sesquiterpenes (Bourtsoukidis et al. 2018).

2.4.3.2 *Historical changes of BVOCs and contribution to climate change*

Climate warming over the past 30 years, together with the longer growing season experienced in boreal and temperate environments, have increased BVOC global emissions by since the preindustrial times (*limited evidence, medium agreement*) (Peñuelas 2009; Sanderson et al. 2003; Pacifico et al. 2012). This was opposed by lower BVOC emissions caused by the historical conversion of natural vegetation and forests to cropland (*limited evidence, medium agreement*) (Unger 2013, 2014a; Fu and Liao 2014). The consequences of historical anthropogenic land cover change were a decrease in the global formation of SOA (-13 %, Scott et al. 2017) and tropospheric burden (-13 %, Heald and Geddes 2016). This has resulted in a positive radiative forcing (and thus warming) from 1850 to 2000 of 0.017 W m^{-2} (Heald and Geddes 2016), 0.025 (Scott et al. 2017) and 0.09 W m^{-2} (Unger 2014b) through the direct aerosol effect. In present-day conditions, global SOA production from all sources spans between 13 and 121 Tg yr^{-1} (Tsigaridis et al. 2014a). The indirect aerosol effect (change in cloud condensation nuclei), resulting from land use induced changes in BVOC emissions, adds an additional positive radiative forcing of 0.008 W m^{-2} (Scott et al. 2017). More studies with different model setups are needed to fully assess this indirect aerosol effect associated with land use change from the preindustrial to present. CMIP6 global emissions pathways (Hoesly et al. 2018; Gidden et al. 2018) estimates global VOCs emissions in 2015 at $230 \text{ Mt VOC yr}^{-1}$. They also estimated that from 2000 to 2015, emissions were up from 200 to $230 \text{ Mt VOC yr}^{-1}$.

There is (*limited evidence, medium agreement*) that historical changes in BVOC emissions have also impacted tropospheric ozone. At most surface locations where land use has changed, the NO_x concentrations are sufficiently high for the decrease in BVOC emissions to lead to decreasing ozone concentrations (Scott et al. 2017). However, in more pristine regions (with low NO_x concentrations), the imposed conversion to agriculture has increased ozone through decreased BVOC emissions and their subsequent decrease in OH (Scott et al. 2017; Heald and Geddes 2016). In parallel, the enhanced soil NO_x emissions from agricultural land, can increase the ozone concentrations in NO_x limited regions (Heald and Geddes 2016).

Another impact of historical decrease in BVOC emissions is the reduction in the atmospheric lifetime of methane (*limited evidence, medium agreement*), which results in a negative radiative forcing that ranges from -0.007 W m^{-2} (Scott et al. 2017) to -0.07 W m^{-2} (Unger 2014b). However, the knowledge of to which degree BVOC emissions impact oxidant concentrations, in particular OH (and thus methane concentrations), is still limited and therefore these numbers are very uncertain (Heald and Spracklen 2015; Scott et al. 2017). The effect of land use change on BVOC emissions are highly heterogeneous (Rosenkranz et al. 2015) and though the global values of forcing described above are small, the local or regional values can be higher and even of opposite sign than the global values.

2.4.3.3 *Future changes of BVOCs*

Studies suggest that increasing temperature will change BVOC emissions through change in species composition and rate of BVOC productions. A further 2°C to 3°C rise in the mean global temperature, could increase BVOC global emissions by an additional 30–45% (Peñuelas and Llusà 2003). In two modelling studies, the impact on climate from rising BVOC emissions have been found to become even larger with decreasing anthropogenic aerosol emissions (Kulmala et al. 2013; Sporre et al. 2018). A negative feedback on temperature, arising from the BVOC-induced increase in the first indirect aerosol effect have been estimated by two studies to be in the order of $-0.01 \text{ W m}^{-2} \text{ K}$ (Scott et al. 2018b; Paasonen et al. 2013). Enhanced aerosol scattering from increasing BVOC emissions has been estimated to contribute with a global gain in BVOC emissions of 7% (Rap et al. 2018). In a warming planet, BVOC emissions are expected to increase but magnitude is unknown and will depend on future land use change, in addition to climate (*limited evidence, medium agreement*).

There is a very limited number of studies investigating the climate impacts of BVOCs using future land use

scenarios (Ashworth et al. 2012; Pacifico et al. 2012). Scott et al. (2018a) found that a future deforestation according to the land use scenario in RCP8.5 leads to a 4% decrease in BVOC emissions at the end of the century. This resulted in a direct aerosol forcing of $+0.006 \text{ W m}^{-2}$ (decreased reflection by particles in the atmosphere) and a first indirect aerosol forcing of -0.001 W m^{-2} (change in the amount of CCN). Studies not including future land use scenarios but investigating the climate feedbacks leading to increasing future BVOC emissions, have found a direct aerosol effect of -0.06 W m^{-2} (Sporre et al. 2018) and an indirect aerosol effect of -0.45 W m^{-2} (Makkonen et al. 2012; Sporre et al. 2018). The stronger aerosol effects from the feedback compared to the land use are, at least partly, explained by a much larger change in the BVOC emissions.

A positive climate feedback could happen in a future scenario with increasing BVOC emissions, where higher ozone and methane concentrations could lead to an enhanced warming which could further increase BVOC emissions (Arneeth et al. 2010). This possible feedback is mediated by NO_x levels. One recent study including dynamic vegetation, land use change, CO₂ and climate change found no increase or even a slight decrease in global BVOC emissions at the end of the century (Hantson et al. 2017). There is a lack of understanding concerning the processes governing the BVOC emissions, the oxidation processes in the atmosphere, the role of the BVOC oxidation products in new particle formation and particle growth, as well as general uncertainties in aerosol-cloud interactions. There is a need for continued research into these processes but the current knowledge indicates that changing BVOC emissions need to be taken into consideration when assessing the future climate and how land use will affect it. In summary, the magnitude and sign of net effect of BVOC emissions on the radiation budget and surface temperature is highly uncertain.

2.5 Land impacts on climate and weather through biophysical and GHGs effects

The focus of this section is summarised **Figure 2.13**. We report on what we know regarding the influence land has on climate via biophysical and biogeochemical exchanges. Biogeochemical effects herein only refer to changes in net emissions of CO₂ from land. The influence of land on atmospheric composition is discussed in Section 2.3.

All sections discuss impacts of land on global and regional climate, and climate extremes, whenever the information is available. Section 2.5.1 presents effects of historical and future land use scenarios; section 2.5.2 is devoted to impacts of specific anthropogenic land uses such as forestation, deforestation, irrigation, crop and forest management; section 2.5.3 focuses on how climate driven land changes feedback on climate and section 2.5.4 puts forward that land use changes in one region can affect another region.

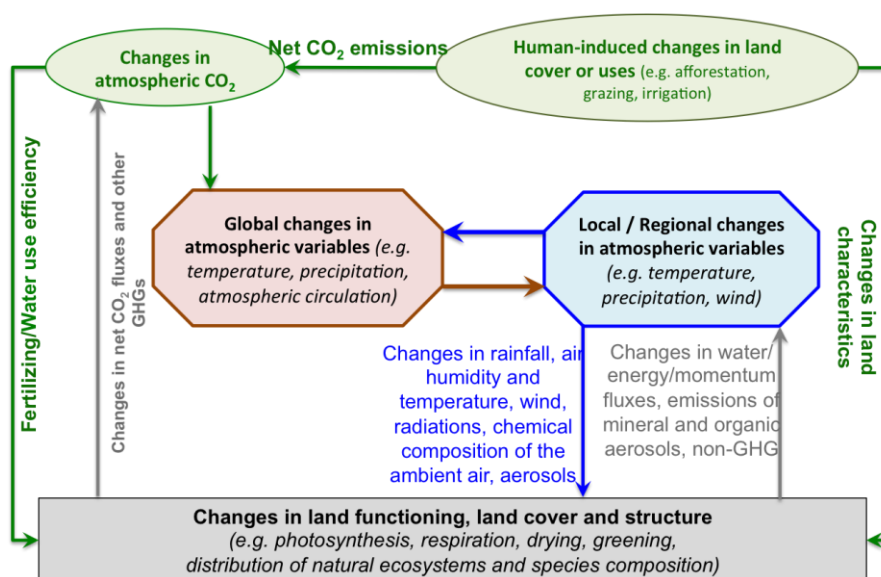


Figure 2.13 Global, local and regional climate changes are the focus of this section. They are examined

through changes in climate states (e.g., changes in air temperature and humidity, rainfall, radiation) as well as through changes in atmospheric dynamics (e.g., circulation patterns). Changes in land that influence climate are either climate- or Human- driven. Green arrows and boxes refer to what we consider herein as imposed changes (forcings). Grey box and arrows refer to responses of land to forcings (green and blue boxes) and feedbacks on those initial forcings. Red and blue boxes and arrows refer respectively to global and local/regional climate changes and their subsequent changes on land

2.5.1 Impacts of historical and future anthropogenic land cover changes

The studies reported below focus essentially on modelling experiments, as there is no direct observation of how historical land use changes have affected the atmospheric dynamics and physics at the global and regional scales. Moreover, the climate modelling experiments only assess the impacts of anthropogenic land cover changes (e.g. deforestation, urbanisation) and neglect the effects of changes in land management (e.g. irrigation, use of fertilisers, choice of species varieties among managed forests or crops). Because of this restricted accounting for land use changes we will use the term land cover changes in the following sub-sections (2.5.1.1 and 2.5.1.2).

Each section starts by describing changes at the global scale, and at the regional scale and ends with what we know about the impacts of those scenarios on extreme weather events, whenever the information is available.

2.5.1.1 Impacts of global historical land cover changes on climate

2.5.1.1.1 At the global level

The contribution of anthropogenic land cover changes to the net global warming throughout the 20th century has been derived from few model-based estimates that account simultaneously for biogeochemical and biophysical effects of land on climate (Table 2.4). The simulated net change in mean global annual surface air temperature, averaged over all the simulations, is a small warming of $0.078 \pm 0.093^\circ\text{C}$, ranging from small cooling simulated by two models (-0.05°C and -0.02°C respectively in (Brovkin et al. 2004) and (Simmons and Matthews 2016), to larger warming simulated by three models ($>+0.14^\circ\text{C}$ (Shevliakova et al. 2013; Pongratz et al. 2010; Matthews et al. 2004). When starting from the Holocene period (He et al. 2014) has estimated an even larger net warming effect of anthropogenic land cover changes ($+0.72^\circ\text{C}$).

Table 2.4 Change in mean global annual surface air temperature resulting from anthropogenic land cover change over the historical period. This historical period varies from one simulation to another (middle column).

Reference of the study	Time period	Mean global annual change in surface air temperature ($^\circ\text{C}$)
(Simmons and Matthews 2016)	1750-2000	-0.02
(Shevliakova et al. 2013)	1861-2005	+0.17
(Pongratz et al. 2010)	1900-2000	+0.14
(Matthews et al. 2004)	1700-2000	+0.15
(Brovkin et al. 2004)	1850-2000	-0.05
Mean \pm standard deviation		0.078 ± 0.093

This net small warming signal results from the competing effects of biophysical cooling (*medium confidence*) and biogeochemical warming (*very high confidence*; Figure 2.14²). The global biophysical cooling alone has been estimated by a larger range of climate models and is $-0.10 \pm 0.14^\circ\text{C}$; it ranges from -0.57°C to $+0.06^\circ\text{C}$ (e.g. (Zhang et al. 2013a; Hua and Chen 2013; Jones et al. 2013b; Simmons and Matthews 2016), Table A2.1). This cooling is essentially dominated by increases in surface albedo: historical land cover changes have generally led to a dominant brightening of land as discussed in AR5 (Myhre et al. 2013). Reduced incoming long-wave radiation at the surface from reduced evapotranspiration and thus less water vapour in the atmosphere has also been reported as a potential contributor to this cooling (Claussen et al. 2001a). The cooling is however dampened by decreases in turbulent fluxes leading to decreased loss of

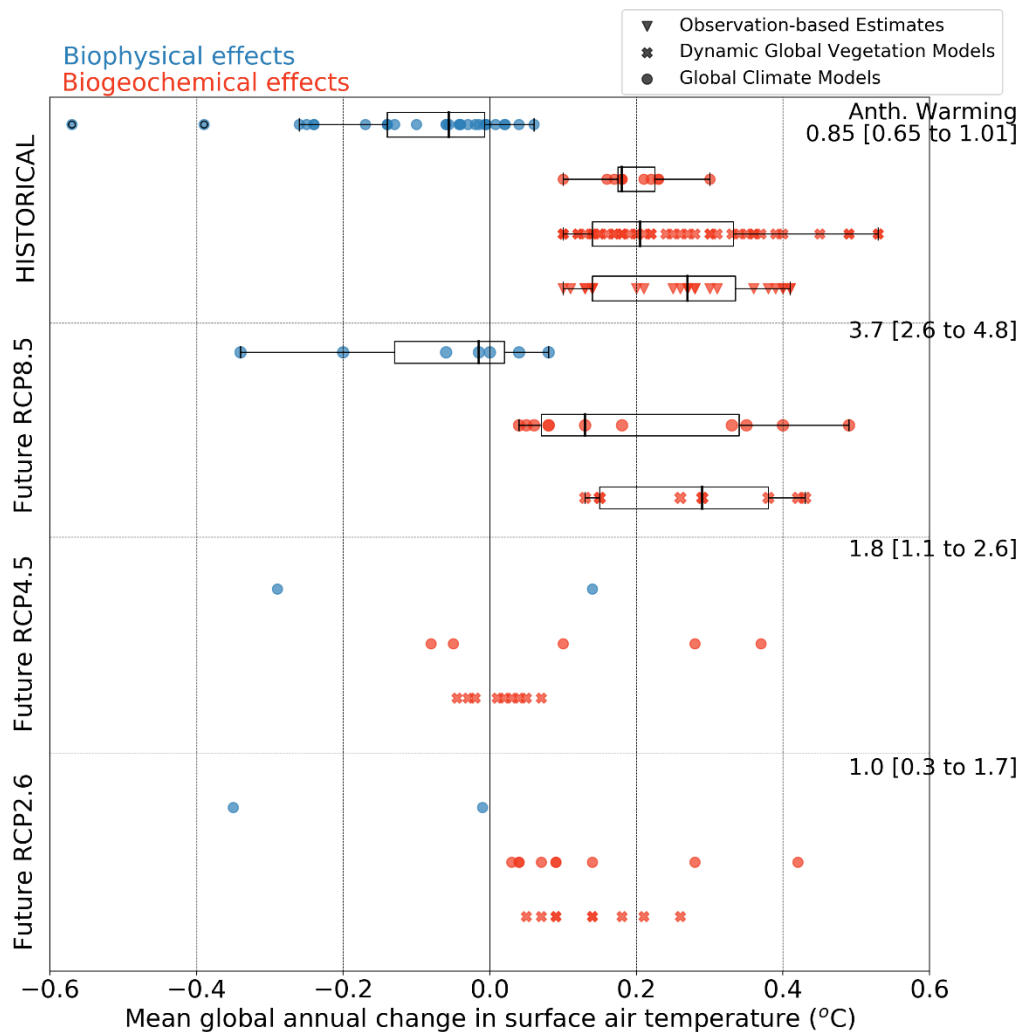
² FOOTNOTE: The detailed list of all values used to construct this figure is provided in Table A2.1 in the Appendix at the end of the Chapter

1 heat and water vapour from the land through convective processes. Those non-radiative processes are indeed
2 well-known to often oppose the albedo-induced surface temperature changes (e.g., (Davin and de Noblet-
3 Ducoudre 2010; Boisier et al. 2012)).

4
5 Historical land cover changes have contributed to the increase in atmospheric CO₂ content (Section 2.3) and
6 thus to global warming (biogeochemical effect, *very high confidence*). The global mean biogeochemical
7 warming has been calculated from observation-based estimates (+0.25±0.10°C; e.g. (Li et al. 2017a;
8 Avitabile et al. 2016; Carvalhais et al. 2014; Le Quéré et al. 2015)), or estimated from dynamic global
9 vegetation models (+0.24±0.12°C; e.g. (Peng et al. 2017; Arneth et al. 2017a; Pugh et al. 2015; Hansis et al.
10 2015)) and global climate models (+0.20±0.05°C; (Pongratz et al. 2010; Brovkin et al. 2004; Matthews et al.
11 2004; Simmons and Matthews 2016)).

12
13 The magnitude of these simulated biogeochemical effects may however be underestimated as they do not
14 account for a number of processes such as land management, nitrogen/phosphorus cycles, changes in the
15 emissions of CH₄, N₂O and non-GHG emissions from land (Ward et al. 2014; Arneth et al. 2017b; Cleveland
16 et al. 2015; Pongratz et al. 2018). Two studies have accounted for those compounds and found a global net
17 positive radiative forcing in response to historical anthropogenic land cover changes, indicating a net surface
18 warming (Mahowald et al. 2017; Ward et al. 2014). However, first the estimated biophysical radiative
19 forcing in those studies only accounts for changes in albedo and not for changes in turbulent fluxes.
20 Secondly, the combined estimates also depend on other several key modelling estimates such as climate
21 sensitivity, CO₂ fertilisation caused by land use emissions, possible synergistic effects, validity of radiative
22 forcing concept for land forcing. The comparison with the other above-mentioned modelling studies is thus
23 difficult.

24
25 In addition, most of those estimates do not account for the evolution of natural vegetation in unmanaged
26 areas, while observations and numerical studies have reported a greening of the land in boreal regions
27 resulting from both extended growing season and poleward migration of tree lines (Lloyd et al. 2003; Lucht
28 et al. 1995), Section 2.2). This greening enhances global warming via a reduction of surface albedo (winter
29 darkening of the land through the snow-albedo feedbacks, e.g. (Forzieri et al. 2017)). At the same time
30 cooling occurs due to increased evapotranspiration during the growing season, along with enhanced
31 photosynthesis, i.e. increased CO₂ sink (Qian et al. 2010). When feedbacks from the poleward migration of
32 treeline is accounted for together with the biophysical effects of historical anthropogenic land cover change,
33 the biophysical annual cooling (about -0.20°C to -0.22°C on land, -0.06°C globally) is significantly
34 dampened by the warming (about +0.13°C) resulting from the movements of natural vegetation (Strengers et
35 al. 2010). Accounting simultaneously for both anthropogenic and natural land cover changes reduces the
36 cooling impacts of historical land cover change in this specific study.
37



1
2 **Figure 2.14** Changes in mean global annual surface air temperature (°C) in response to historical and
3 **future anthropogenic land cover changes** as estimated from a range of studies (see Table A2.1 in the
4 **Appendix for detailed information**). Temperature changes resulting from biophysical processes (e.g.
5 **changes in physical land surface characteristics such as albedo, evapotranspiration, and roughness**
6 **length) are illustrated using blue symbols**; temperature changes resulting from biogeochemical processes
7 **(e.g. changes in atmospheric CO₂ composition) use red symbols**. Future changes are shown for three
8 **distinct scenarios: RCP8.5, RCP4.5 & RCP2.6**. The markers ‘filled circle’, ‘filled cross’, and ‘filled
9 **triangle down**’ represent estimates from respectively **global climate models, dynamic global vegetation**
10 **models (DGVMs), and observations**. When the number of estimate is sufficiently large, box plots are
11 **overlaid; they show the ensemble minimum, first quartile (25th percentile), median, third quartile (75th**
12 **percentile), and the ensemble maximum**. Scatter points beyond the box plot are the outliers. Details
13 **about how temperature change is estimated from DGVMs and observations is provided in Appendix**.
14 **Numbers on the right hand-side give the mean and the range of simulated mean global annual warming**
15 **from various climate models**.
16

17 **2.5.1.1.2 At the regional level**

18 The global and annual estimates reported above mask out very contrasted regional and seasonal differences.
19 Biogeochemical effects of anthropogenic land cover change on temperature follow the spatial patterns of
20 GHG-driven climate change with stronger warming over land than ocean, and stronger warming in northern
21 high latitudes than in the tropics and equatorial regions (Arctic amplification). Biophysical effects on the
22 contrary are much stronger where land cover has been modified than in their surroundings (see 2.5.4 for a
23 discussion on non-local effects). Very contrasted regional temperature changes can thus result depending on
24 whether biophysical processes dampen or exacerbate biogeochemical impacts.
25

Figure 2.15 compares, for seven climate models, the biophysical effects of historical anthropogenic land cover change in North America and Eurasia (essentially cooling) to the regional warming resulting from the increased atmospheric CO₂ content since pre-industrial times ((De Noblet-Ducoudré et al. 2012); comparing 1973–2002 to 1871–1900). It shows a dominant biophysical cooling effect of changes in land cover, at all seasons, as large as the regional footprint of anthropogenic global warming. Averaged over all agricultural areas of the world (Pongratz et al. 2010) reported a 20th century biophysical cooling of -0.10°C, and (Strengers et al. 2010) reported a land induced cooling as large as -1.5°C in western Russia and eastern China between 1871 and 2007. There is thus *medium confidence* that anthropogenic land cover change has dampened warming in many regions of the world over the historical period.

Very few studies have explored the effects of historical land cover changes on seasonal climate. There are however evidences that the seasonal magnitude and sign of those effects at the regional level are strongly related to soil-moisture/evapotranspiration and snow regimes, particularly in temperate and boreal latitudes (Teuling et al. 2010; Pitman and de Noblet-Ducoudré 2012; Alkama and Cescatti 2016). Quesada et al. (2017a) showed that atmospheric circulation changes can be significantly strengthened in winter for tropical and temperate regions. However, the lack of studies underlines the need for a more systematic assessment of seasonal, regional and other than mean temperature metrics in the future.

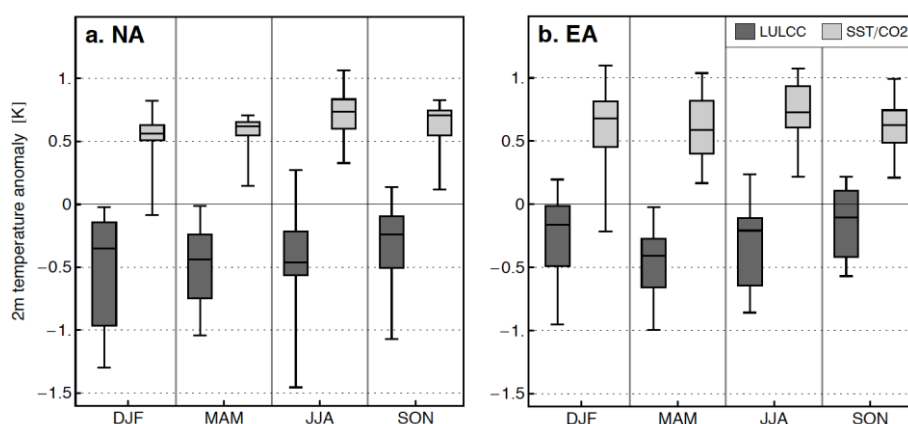


Figure 2.15 Simulated changes in mean surface air temperature (°C) between the pre-industrial period (1870–1900) and present-day (1972–2002) for all seasons and for a) North America and b) Eurasia (De Noblet-Ducoudré et al. 2012). Light grey boxes are the changes simulated in response to increased atmospheric GHG content between both time periods and subsequent changes in sea-surface temperature and sea-ice extent (SST/CO₂); the CO₂ changes accounted for include emissions from all sources including land use. Dark grey boxes are the changes simulated in response to the biophysical effects of historical land cover changes. The box-and-whisker plots have been drawn using results from seven climate models and ensembles of ten simulations per model and time period. The bottom and top of the each grey box are the 25th and 75th percentiles, and the horizontal line within each box is the 50th percentile (the median). The whiskers (straight lines) indicate the ensemble maximum and minimum values. Seasons are respectively December-January-February (DJF), March-April-May (MAM), June-July-August (JJA) and September-October-November (SON). North America and Eurasia are extended regions where land use changes are the largest between the two time periods considered (their contours can be found in Figure 1 of (De Noblet-Ducoudré et al. 2012)).

2.5.1.1.3 Effects on extremes

The effect of historical deforestation on extreme temperature trends is intertwined with the effect of other climate forcings thus making it difficult to quantify based on observations. Based on results from four climate models, the impact of historical anthropogenic land cover change on temperature and precipitation extremes was found to be locally as important as changes arising from increases in atmospheric CO₂ and sea-surface temperatures, but with a lack of model agreement on the sign of changes (Pitman et al. 2012). In some regions the impact of land cover change masks or amplifies the effect of increased CO₂ on extremes (Avila et al. 2012; Christidis et al. 2013). Using an observational constraint for the local biophysical effect of land cover change applied to a set of CMIP5 climate models, (Lejeune et al. 2018) found that historical deforestation increased extreme hot temperatures in northern mid-latitudes. The results also indicate a stronger impact on the warmest temperatures compared to mean temperatures. Findell et al. (2017) reached similar conclusions, although using only a single climate model. Importantly, the climate models involved in

1 these three studies did not consider the effect of management changes which have been shown to be
 2 important, as discussed Section 2.5.2.

3
 4 Based on the studies discussed above there is yet *limited evidence* but *high agreement* that land cover change
 5 affects local temperature extremes more than mean values. Observational studies assessing the role of land
 6 cover on temperature extremes are still very limited (Zaitchik et al. 2006; Renaud and Rebetez 2008), but
 7 suggest that trees dampen seasonal and diurnal temperature variations at all latitudes and even more so in
 8 temperate regions compared to short vegetation (Chen et al. 2018; Duveiller et al. 2018b; Li et al. 2015a; Lee
 9 et al. 2011). Furthermore, trees also locally dampen the amplitude of heat extremes (Renaud and Rebetez
 10 2008; Zaitchik et al. 2006) although this result depends on the forest type, coniferous trees providing less
 11 cooling effect than broadleaf trees (Renaud et al. 2011; Renaud and Rebetez 2008).

12 13 **2.5.1.2 Impacts of future global land cover changes on climate**

14 **2.5.1.2.1 At the global level**

15 The most extreme emissions scenario (RCP8.5) that has been developed for the last coordinated modelling
 16 intercomparison of climate models (CMIP5) is the one that has received the most attention in the literature
 17 with respect to how projected future anthropogenic land cover changes (Hurtt et al. 2011) will affect the very
 18 large simulated global warming.

19
 20 Seven model-based studies have examined both the biophysical and biogeochemical effects of anthropogenic
 21 changes in land-cover, as projected in RCP8.5, on future climate change (Table 2.5 ; (Simmons and
 22 Matthews 2016; Davies-Barnard et al. 2014; Boysen et al. 2014)). They all agree on a biogeochemical
 23 warming, ranging from +0.04°C to +0.35°C, in response to land cover change. Two models predict an
 24 additional biophysical warming, while the others agree on a biophysical cooling that dampens (or overrules)
 25 the biogeochemical warming. Using a wider range of global climate models, the biogeochemical warming
 26 (*high confidence*) is +0.20±0.15°C whereas it is +0.28±0.11°C when estimated from dynamic global
 27 vegetation models (Pugh et al. 2015; Stocker et al. 2014). This biogeochemical warming is compensated for
 28 by a biophysical cooling (*medium confidence*) of -0.10±0.14°C (Quesada et al. 2017a; Davies-Barnard et al.
 29 2015; Boysen et al. 2014). The estimates of temperature changes resulting from anthropogenic land cover
 30 changes alone remain very small compared to the projected mean warming of +3.7°C by the end of the 21st
 31 century (ranging from 2.6 to 4.8°C depending on the model and compared to 1986-2005; Figure 2.14).

32
 33 **Table 2.5 Change in mean global annual surface air temperature resulting from**
 34 **anthropogenic land cover changes projected for the future, according to three different**
 35 **scenarios: RCP8.5, RCP4.5 and RCP2.6. Temperature change resulting from biophysical**
 36 **and biogeochemical effects of land cover change are examined.**

Reference of the study	Time period	Mean global annual change in surface air temperature (°C)		
		Biophysical / Biogeochemical		
		RCP2.6	RCP4.5	RCP8.5
(Simmons and Matthews 2016)	2000-2100	-0.35 / +0.42	-0.29 / + 0.37	-0.34 / + 0.35
(Davies-Barnard et al. 2014)	2005-2100	-0.01 / +0.04	+0.14 / -0.08	-0.015 / +0.04
(Boysen et al. 2014)	2005-2100			+0.04 / +0.08 0 / +0.05 +0.08 / +0.06 -0.20 / +0.13 -0.06 / +0.33

37
 38 Two other projected land cover change scenarios have been examined (RCP4.5 and RCP2.6; Table 2.5 ;
 39 Figure 2.14) but only one climate modelling experiment has been carried out for each, to estimate the
 40 biophysical impacts on climate of those changes (Davies-Barnard et al. 2015). For RCP2.6, earth system and
 41 dynamic global vegetation models agree on a systematic biogeochemical warming resulting from the
 42 imposed land cover changes, ranging from +0.03 to +0.28°C (Brovkin et al. 2013a), which is significant
 43 compared to the projected mean climate warming of +1°C by the end of the 21st century (ranging from 0.3 to
 44 1.7°C depending on the models, compared to 1986-2005). A very small biophysical cooling is expected from
 45 the one estimate. For RCP4.5 biophysical warming is expected from only one estimate, and results from a

1 projected large forestation in the temperate and high latitudes. There is no agreement on the sign of the
2 biogeochemical effect: there are as many studies predicting cooling as warming, whichever the method to
3 compute those effects (earth system models or dynamic global vegetation models).
4

5 Previous scenarios (Special Report on Emission Scenarios (SRES), results of climate studies using those
6 scenarios were reported in AR4) displayed larger land use changes than the more recent ones (RCP, AR5).
7 There is *low confidence* from some of those previous scenarios (SRES A2 and B1) of a small warming effect
8 (+0.2 to +0.3°C) of anthropogenic land cover change on mean global climate, this being dominated by the
9 release of CO₂ in the atmosphere from land conversions (Sitch et al. 2005). This additional warming remains
10 quite small when compared to the one resulting from the combined anthropogenic influences [+1.7°C for
11 SRES B1 and +2.7°C for SRES A2]. A global biophysical cooling of -0.14°C is estimated in response to the
12 extreme land cover change projected in SRES A2, a value that far exceeds the impacts of historical land use
13 changes (-0.05°C) calculated using the same climate model (Davin et al. 2007). The authors derived a
14 biophysical climatic sensitivity to land use change of about -0.3°C W.m⁻² for their model, whereas a warming
15 of about 1°C W.m⁻² is obtained in response to changes in atmospheric CO₂ concentration.
16

17 Those studies generally do not report on changes in atmospheric variables other than surface air temperature,
18 thereby limiting our ability to assess the effects of anthropogenic land cover changes on regional climate
19 Sitch et al. (2005). However, reported small reductions in rainfall via changes in biophysical properties of
20 the land, following the massive tropical deforestation in SRES A2 (+0.5 and +0.25 mm day⁻¹ respectively in
21 the Amazon and Central Africa). They also report opposite changes, that is increased rainfall of about 0.25
22 mm day⁻¹ across the entire tropics and subtropics, triggered by biogeochemical effects of this same
23 deforestation.
24

25 **2.5.1.2.2 At the regional level**

26 In regions that will undergo land cover changes, dampening of the future anthropogenic warming can be as
27 large as -26% while enhancement is always smaller than 9% within RCP8.5 by the end of the 21st century
28 (Boysen et al. 2014). Voltaire (2006) show that, by 2050 and following the SRES B2 scenario, the
29 contribution of land cover changes to the total temperature change can be as large as 15% in many boreal
30 regions, and as large as 40% in south western tropical Africa. Feddema et al. (2005) simulate large decreases
31 in the diurnal temperature range in the future (2050 and 2100 in SRES B1 and A2) following tropical
32 deforestation in both scenarios. In the Amazon for example the diurnal temperature range is lowered by
33 2.5°C due to increases in minimum temperature while little change is obtained for the maximum value.

34 There is thus *medium evidence* that future anthropogenic land cover change will have a significant effect on
35 regional temperature via biophysical effects in many regions of the world. There is however *no agreement*
36 on whether warming will be dampened or enhanced and there is *no agreement* on the sign of the contribution
37 across regions.
38

39 There are very few studies that go beyond analysing the changes in mean surface air temperature. Some
40 studies attempted to look at global changes in rainfall and found no significant influence of future land cover
41 changes (Brovkin et al. 2013a; Sitch et al. 2005; Feddema et al. 2005). Quesada et al. (2017a,b) however
42 carried out a systematic multi-model analysis of the response of a number of atmospheric, radiative and
43 hydrological variables (e.g. rainfall, sea level pressure, geopotential height, wind speed, soil-moisture,
44 turbulent heat fluxes, shortwave and longwave radiation, cloudiness) to RCP8.5 land cover scenario. In
45 particular, they found a significant reduction of rainfall in 6 out of 8 monsoon regions studied (Figure 2.16)
46 of about 1.9% to 3% (which means more than -0.5mm day⁻¹ in some areas) in response to future
47 anthropogenic land cover changes. Including those changes in global climate models reduces the projected
48 increase in rainfall by about 9% to 41% in those same regions, when all anthropogenic forcings are
49 accounted for (30% in the Global Monsoon region as defined by (Wang and Ding 2008)). In addition, they
50 found a shortening of the monsoon season of one to four days. They conclude that the projected future
51 increase in monsoon rains may be overestimated by those models that do not yet include biophysical effects
52 of land cover changes. Overall, the regional hydrological cycle was found to be substantially reduced and
53 wind speed significantly strengthened in response to regional deforestation within the tropics, with
54 magnitude comparable to projected changes with all forcings (Quesada et al. 2017b).
55

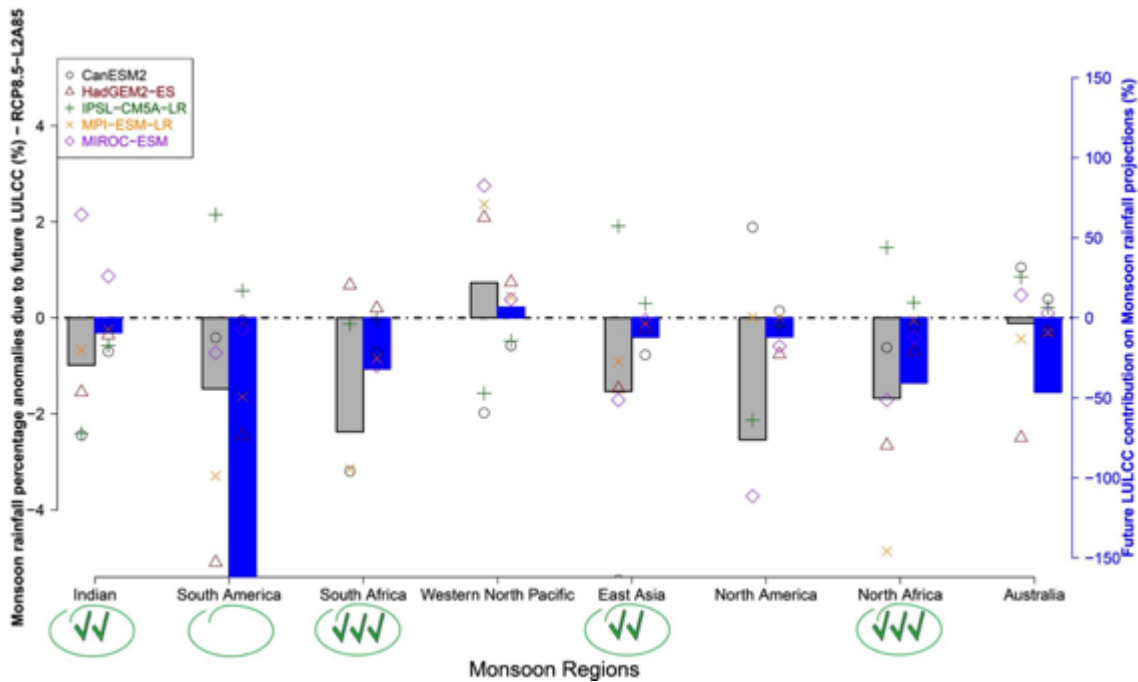


Figure 2.16 Changes in monsoon rainfall in RCP8.5 scenario resulting from projected changes in anthropogenic land cover, in eight monsoonal regions (%), grey bars). Differences are calculated between the end of the 21st century (2071–2100) and the end of the 20th century (1976–2005); percent change is calculated with reference to 1976–2005. Blue bars refer to the relative contribution of land cover changes (in %) to future rainfall projections: it is the ratio between the change in rainfall responding to land cover changes and the one responding to all anthropogenic changes (Quesada et al. 2017b). Negative values mean that changes in land cover have an opposite effect (dampening) on rainfall compared to the effects of all anthropogenic changes. Monsoon regions have been defined following (Yim et al. 2014). The changes have been simulated by five climate models (Brovkin et al. 2013, symbols). Results are shown for December-January-February for southern hemisphere regions, and for June-July-August for northern hemisphere regions. Statistical significance is given by green tick marks and circles: one, two, and three green tick marks are displayed for the regions where at least 80%, 75%, and 80th confidence level, respectively; green circles are added when the regional values are also significant at 90th confidence level. Note that future land cover change impacts on South American monsoon are neither significant nor robust among models, along with very small future projected changes in South American monsoon rainfall.

2.5.1.2.3 Effects on extremes

Results from a set of climate models have shown that the impact of future anthropogenic land cover change on extreme temperatures can be of similar magnitude as the changes arising from half a degree global mean annual surface temperature change (Hirsch et al. 2018). However, this study also found a lack of agreement between models with respect to the magnitude and sign of changes, thus making land cover change a factor of uncertainty in future climate projections.

2.5.2 Impacts of specific land use changes

2.5.2.1 Impacts of deforestation and forestation

Deforestation or forestation³, wherever it occurs, triggers simultaneously warming and cooling of the surface and of the atmosphere via changes in its various characteristics (Pitman 2003; Strengers et al. 2010; Bonan 2008b). Following deforestation, warming results from a) the release of CO₂ and other GHG in the atmosphere (biogeochemical impact) and subsequent increase in incoming infrared radiation at surface (greenhouse effect), b) a decreased in the total loss of energy through turbulent fluxes (latent and sensible

³ FOOTNOTE: The term « forestation » is used herein as this chapter does not distinguish between afforestation and reforestation. In model-based studies, simulations with and without trees are compared; in observation-based estimates, sites with and without trees are compared.

1 heat fluxes) resulting from reduced surface roughness, c) an increased incoming solar radiation following
2 reduced cloudiness that often (but not always) accompanies the decreased total evapotranspiration. Cooling
3 occurs in response to d) increased surface albedo that reduces the amount of absorbed solar radiation, e)
4 reduced incoming infrared radiation triggered by the decreased evapotranspiration and subsequent decrease
5 in atmospheric water vapour. Points b-c-d-e are referred to as biophysical effects. Deforestation and
6 forestation also alter rainfall and winds (horizontal as well as vertical as will be further discussed below).
7

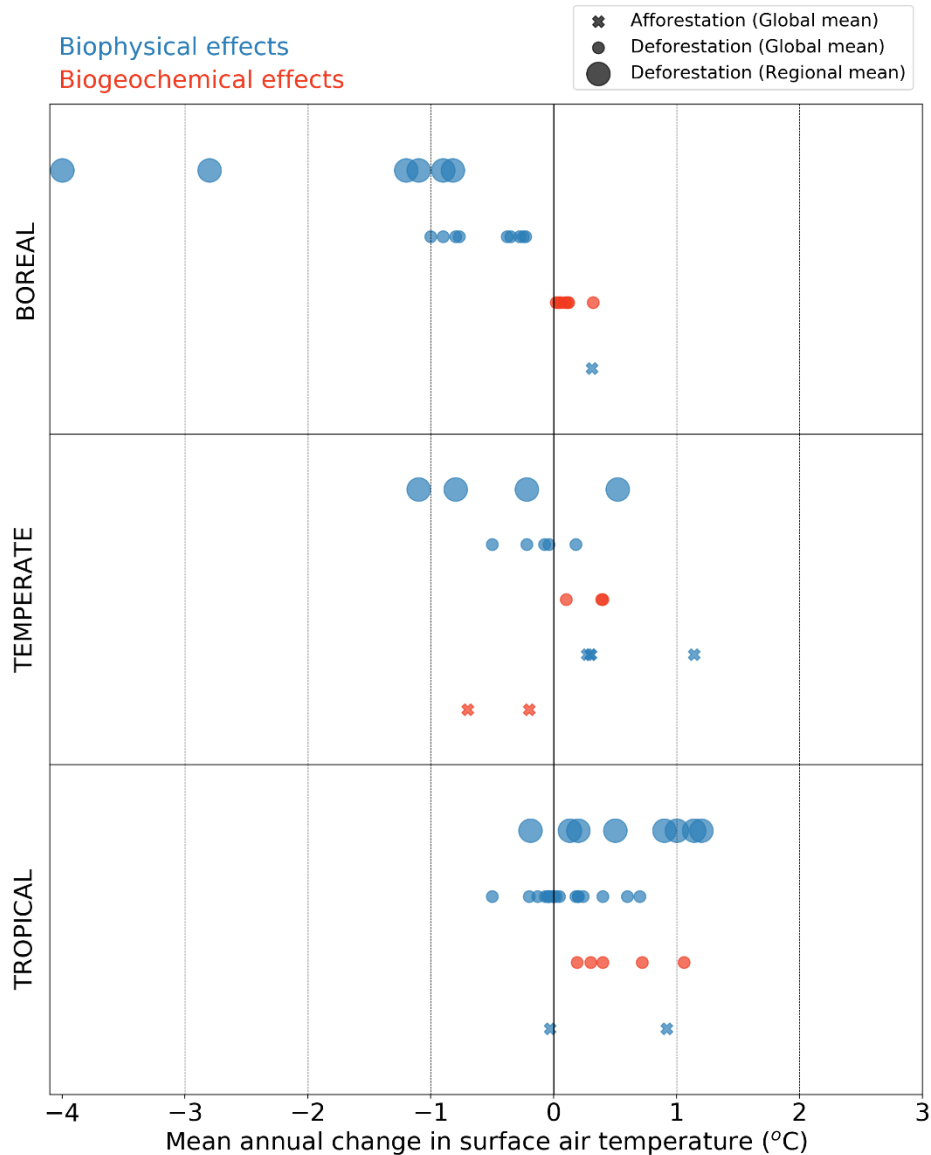
8 The literature that discusses the effects of forestation on climate is more limited than for deforestation, but
9 they reveal a similar climatic response with opposite sign as further discussed below. For each latitudinal
10 band (tropical, temperate and boreal) how very large scale deforestation or forestation impacts global mean
11 climate is examined, followed by examination of the large-scale changes in the specific latitudinal band and
12 end up with more regionally focused analysis. Large scale idealised deforestation or forestation experiments
13 are often carried out with global or regional climate models as they allow to understand and measure how
14 sensitive climate is to very large changes in land cover (similar to the instant doubling of CO₂ in climate
15 models to calculate the climatic sensitivity to GHGs). Details of the model-based studies discussed below
16 can be found in Table A2.2 in the Appendix.
17

18 2.5.2.1.1 *Global and regional impacts of deforestation/forestation in tropical regions*

19 A pan-tropical deforestation would lead to the net release of CO₂ from land and thus to mean global annual
20 warming, with model-based estimates of biogeochemical effects ranging from +0.19 to +1.06°C, with a
21 mean value of +0.53±0.32°C (Ganopolski et al. 2001; Snyder et al. 2004; Devaraju et al. 2015a; Longobardi
22 et al. 2016; Perugini et al. 2017). There is however *no agreement* between models on the magnitude and sign
23 of the biophysical effect of such changes at the global scale (the range spans from -0.5°C to +0.7°C with a
24 mean value of +0.1±0.27°C; Figure 2.17 ; e.g. (Devaraju et al. 2015c; Snyder 2010; Longobardi et al.
25 2016a)). This is the result of many compensation effects in action: increased surface albedo following
26 deforestation, decreased atmospheric water vapour content due to less tropical evapotranspiration, decreased
27 loss of energy from tropical land in the form of latent and sensible heat fluxes.
28

29 There is however *high confidence* that such large land cover change would lead to a mean biophysical
30 warming when averaged over the deforested land. A mean warming of +0.61±0.48°C is found over the entire
31 tropics. Reversely, biophysical regional cooling and global warming is expected from forestation (Wang et
32 al. 2014b; Bathiany et al. 2010a).
33

34 Large-scale deforestation (whether pan-tropical or imposed at the sub-continent level, e.g. the Amazon)
35 results in significant mean rainfall decrease (Lawrence and Vandecar 2015; Lejeune et al. 2015; Perugini et
36 al. 2017). In their review (Perugini et al. 2017) reported an average simulated decrease of -288 ± 75 mm yr⁻¹
37 (95%-confidence interval). Inversely large-scale forestation increases tropical rainfall by 41 ± 21 mm yr⁻¹.
38 The magnitude of the change in precipitation strongly depends on the type of land cover conversion. For
39 instance, conversion of tropical forest to bare soil causes larger reductions in regional precipitation than
40 conversion to pasture (respectively -470 ± 60 mm yr⁻¹ and -220 ± 100 mm yr⁻¹). Biogeochemical effects in
41 response to pan-tropical deforestation, particularly CO₂ release, are generally not taken into account in those
42 studies but could intensify the hydrological cycle and thus precipitation (Kendra Gotangco Castillo and
43 Gurney 2013).
44



1
 2 **Figure 2.17 Changes in mean annual surface air temperature (°C) in response to idealised large scale**
 3 **deforestation (circles) or forestation (crosses), estimated from a range of studies (see Table A2.2 in the**
 4 **Appendix for detailed information and references to the studies). Temperature changes resulting from**
 5 **biophysical processes (e.g. changes in physical land surface characteristics such as albedo,**
 6 **evapotranspiration, and roughness length) are illustrated using blue symbols; temperature changes**
 7 **resulting from biogeochemical processes (e.g. changes in atmospheric CO₂ composition) use red**
 8 **symbols. Small blue and red circles, and crosses, are model-based estimates of changes in temperature**
 9 **averaged globally. Large circles are estimates averaged only over the latitudinal band where**
 10 **deforestation is imposed.**
 11

12 Specific model-based deforestation studies have been carried out for Africa (Hagos et al. 2014; Boone et al.
 13 2016; Xue et al. 2016; Nogherotto et al. 2013; Hartley et al. 2016; Klein et al. 2017; Abiodun et al. 2012),
 14 southern America (Butt et al. 2011; Wu et al. 2017; Spracklen and Garcia-Carreras 2015; Lejeune et al.
 15 2015), South-East Asia (Ma et al. 2013b; Werth and Avissar 2005; Mabuchi et al. 2005; Tölle et al. 2017).
 16 All found decreases in evapotranspiration following deforestation (*high agreement*), resulting in surface
 17 warming despite the competing effect from increased surface albedo (*high agreement*). Changes in thermal
 18 gradients between deforested and adjacent regions, between land and ocean, affect horizontal surface winds
 19 (*high agreement*) and thus modify the areas where rainfalls as discussed in Section 2.5.4. An increase in the
 20 land-sea thermal contrast has been found in many studies as surface friction is reduced by deforestation, thus

1 increasing the monsoon flow in Africa and South America (Wu et al. 2017).

2
3 Observation-based estimates all agree that deforestation increases local land-surface and ambient air
4 temperatures in the tropics, while forestation has the reverse effect (*very high confidence*; (Prevedello et al.
5 2019; Schultz et al. 2017; Li et al. 2015b; Alkama and Cescatti 2016)). There is *very high confidence* that
6 forests are cooler than any shorter vegetation (crops, grasses, bare soil) during daytime due to larger
7 transpiration rates, and there is *high confidence* that the amplitude of the diurnal cycle is smaller in the
8 presence of forests.

9
10 Large-scale forestation scenarios of West Africa (Abiodun et al. 2012), eastern China (Ma et al. 2013a) or
11 Saharan and Australian deserts (Ornstein et al. 2009; Kemena et al. 2017) all concluded that regional surface
12 cooling is simulated wherever trees are grown (-2.5°C in the Sahel and -1°C in the Savanna area of West
13 Africa, up to -8°C in western Sahara, -1.21°C over land in eastern China) while cooling of the ambient air is
14 smaller (-0.16°C). In the case of Savanna forestation this decrease entirely compensates the GHG induced
15 future warming ($+1^{\circ}\text{C}$ following the SRES A1B scenario). West African countries thus have the potential to
16 reduce, or even totally cancel at some places, the GHG-induced warming in the deforested regions (Abiodun
17 et al. 2012). However, this is compensated by enhanced warming in adjacent countries (non-local effect).

18 19 **2.5.2.1.2 Global and regional impacts of deforestation/forestation in temperate regions**

20 As for the tropics, model-based experiments show that large-scale temperate deforestation would induce a
21 small mean global annual warming through the net release of CO_2 into the atmosphere (ranging from $+0.10$
22 to $+0.40^{\circ}\text{C}$ with a mean value of $+0.20 \pm 0.13^{\circ}\text{C}$, Figure 2.17), whereas there is less agreement on the sign of
23 the mean global annual temperature change resulting from biophysical processes: estimates range from -
24 0.5°C to $+0.18^{\circ}\text{C}$ with a mean value of $-0.13 \pm 0.22^{\circ}\text{C}$. There is also *very low agreement* on the mean annual
25 temperature change in the temperate zone ($-0.4 \pm 0.62^{\circ}\text{C}$; (Phillips et al. 2007; Snyder et al. 2004b;
26 Longobardi et al. 2016a; Devaraju et al. 2015a, 2018b)). There is *medium agreement* on a global and
27 latitudinal biophysical warming in response to forestation (Figure 2.17 ; (Laguë and Swann 2016a; Swann et
28 al. 2012a; Gibbard et al. 2005; Wang et al. 2014b)), but this is based on a smaller number of studies.

29
30 The lack of agreement at the annual scale among the climate models is however masking *rising agreement*
31 regarding seasonal impacts of deforestation at those latitudes. There is *high agreement* that temperate
32 deforestation leads to summer warming and winter cooling (Bright et al. 2017; Zhao and Jackson 2014;
33 Gálos et al. 2011, 2013; Wickham et al. 2013; Ahlswede and Thomas 2017; Anderson-Teixeira et al. 2012;
34 Anderson et al. 2011; Chen et al. 2012; Strandberg and Kjellström 2018). The winter cooling is driven by the
35 increased surface albedo, amplified by the snow-albedo feedback. In some models and when deforestation is
36 simulated for very large areas, the cooling is further amplified by high latitude changes in sea-ice and snow
37 extent (polar amplification). Summer warming occurs because the latent and sensible heat fluxes, that take
38 energy out of the surface, diminish with the smaller roughness length and lower evapotranspiration
39 efficiency of low vegetation as compared to tree canopies (Davin and de Noblet-Ducoudre 2010; Anav et al.
40 2010). Conversely, there is *high agreement* that forestation in North America or in Europe cools surface
41 climate during summer time, especially in regions where water availability can support large
42 evapotranspiration rates. In temperate regions with water deficits, the simulated change in evapotranspiration
43 following forestation will be insignificant while the decreased surface albedo will favour surface warming.

44
45 Observation-based estimates confirm the existence of a seasonal pattern of response to deforestation, with
46 colder winters anytime there is snow on the ground and anywhere soils are brighter than trees, and warmer
47 summers (Schultz et al. 2017; Wickham et al. 2014; Juang et al. 2007; Tang et al. 2018; Peng et al. 2014;
48 Zhang et al. 2014b; Prevedello et al. 2019; Li et al. 2015b; Alkama and Cescatti 2016). In contrast
49 forestation induces cooler summers wherever trees have access to sufficient soil moisture to transpire. The
50 magnitude of the cooling depends on the wetness of the area of concern (Wickham et al. 2013) as well as on
51 the original and targeted species and varieties implicated in the vegetation conversion (Peng et al. 2014;
52 Juang et al. 2007).

53
54 There is also *high confidence* from observation-based estimates that mean annual daytime temperatures are
55 warmer following deforestation, while night time temperatures are cooler (Schultz et al. 2017; Wickham et
56 al. 2014; Juang et al. 2007; Tang et al. 2018; Prevedello et al. 2019; Peng et al. 2014; Zhang et al. 2014b; Li

1 et al. 2015b; Alkama and Cescatti 2016). Deforestation then increases the amplitude of diurnal temperature
2 variations while forestation reduces it (*high confidence*). Two main reasons have been put forward to explain
3 why nights are warmer in forested areas: their larger capacity to store heat, and the existence of a nocturnal
4 temperature inversion bringing warmer air from aloft.

5
6 In addition to those seasonal and diurnal fluctuations, (Lejeune et al. 2018) found systematic warming of the
7 hottest summer days following historical deforestation in the northern mid-latitudes, and this echoes
8 (Strandberg and Kjellström 2018) who argue that the August 2003 and July 2010 heat-waves could have
9 been largely mitigated if Europe had been largely forested.

10
11 In a combined modelling of large-scale forestation of western Europe and climate change scenario (SRES
12 A2) (Gálos et al. 2013) found a relatively small dampening potential of additional forest on ambient air
13 temperature at the end of the 21st century when compared to the beginning (the cooling resulting from land
14 cover changes is -0.5°C whereas the GHG-induced warming exceeds 2.5°C). Influence on rainfall was
15 however much larger and significant. Projected annual rainfall decreases following warming were cancelled
16 in Germany and significantly reduced in both France and Ukraine through forestation. In addition forestation
17 also decreased the number of warming-induced dry days, but increased the number of extreme precipitation
18 events.

19
20 The net impact of forestation, combining both biophysical and biogeochemical effects, has been tested in the
21 warmer world predicted by RCP 8.5 scenario (Sonntag et al. 2016, 2018). The cooling effect from the
22 addition of 8 Mkm^2 of forests following the land use RCP 4.5 scenario was too small (-0.27°C annually) to
23 dampen the RCP 8.5 warming. It however reached about -1°C in some temperate regions and -2.5°C in
24 boreal ones. This is accompanied by a reduction in the number of extremely warm days.

25 26 **2.5.2.1.3 Global and regional impacts of deforestation/forestation in boreal regions**

27 Consistent with what we have previously discussed for temperate and tropical regions, large-scale boreal
28 deforestation induces a biogeochemical warming of $+0.11 \pm 0.09^{\circ}\text{C}$ (Figure 2.17). But contrary to those other
29 latitudinal bands, the biophysical effect is a consistent cooling across all models ($-0.55 \pm 0.29^{\circ}\text{C}$ when
30 averaged globally). It is also significantly larger than the biogeochemical warming (e.g. (Dass et al. 2013;
31 Longobardi et al. 2016a; Devaraju et al. 2015a; Bathiany et al. 2010a; Devaraju et al. 2018b)). It is driven by
32 the increased albedo, enhanced by the snow-albedo feedback as well as by an increase in sea-ice extent in the
33 Arctic. Over the boreal lands, the cooling is as large as $-1.8 \pm 1.2^{\circ}\text{C}$. This mean annual cooling however
34 masks out a seasonal contrast as discussed in (Strandberg and Kjellström 2018) and (Gao et al. 2014): during
35 summer time, following the removal of forest, the decreased evapotranspiration results in a significant
36 summer warming that outweighs the effect of an increased albedo effect.

37
38 The same observation-based estimates as discussed in the previous sub-section show similar patterns as for
39 the temperate latitudes: seasonal and daily contrasts. (Schultz et al. 2017) however found that mean annual
40 nighttime changes are as large as daytime ones in those regions (mean annual nocturnal cooling $-1.4 \pm 0.10^{\circ}\text{C}$,
41 balanced by mean annual daytime warming of $1.4 \pm 0.04^{\circ}\text{C}$). This contrasts with both temperate and tropical
42 regions where daytime changes are always larger than nighttime ones.

43
44 Arora and Montenegro (2011) combined large-scale forestation and climate change scenario (SRES A2):
45 forestation of either 50% or 100% of the total agricultural area was gradually prescribed between years 2011
46 and 2060 everywhere. In addition, boreal, temperate and tropical forestation have been tested separately.
47 Both biophysical and biogeochemical effects were accounted for. The net simulated impact of forestation
48 was a cooling varying from -0.04°C to -0.45°C , depending on the location and magnitude of the additional
49 forest cover. It was, however, quite marginal compared to the large global warming resulting from
50 anthropogenic GHG emissions ($+3^{\circ}\text{C}$ at the end of the 21st century). In their experiment, forestation in
51 boreal regions led to biophysical warming and biogeochemical cooling that compensated each other, whereas
52 forestation in the tropics led to both biophysical and biogeochemical cooling. The authors concluded that
53 tropical forestation is three times more effective in cooling down climate than are boreal or temperate
54 forestation.

2.5.2.1.4 Conclusion

In conclusion, planting trees will always result in capturing more atmospheric CO₂ and thus in mean annual cooling of the globe (*very high confidence*). At the regional level however the magnitude and sign of the local temperature change depends on a) where forestation occurs, b) its magnitude, c) the level of warming under which the land cover change is applied and d) the land conversion type. This is because the background climatic conditions (e.g., precipitation and snow regimes, mean annual temperature) within which the land cover changes occur vary across regions (Pitman et al. 2011; Montenegro et al. 2009; Juang et al. 2007; Wickham et al. 2014; Hagos et al. 2014; Voltaire 2006; Feddema et al. 2005; Strandberg and Kjellström 2018). In addition there is *high confidence* that estimates of the influence of any land cover or land use change on surface temperature from the sole consideration of the albedo and the CO₂ effects is incorrect as changes in turbulent fluxes (i.e., latent and sensible heat fluxes) are large contributors to local temperature change (Bright et al. 2017).

There is *high confidence* that in boreal and temperate latitudes the presence of forest cools temperature in warmer locations and seasons provided that the soil is not dry, whereas it warms temperature in colder locations and seasons provided the soil is brighter than trees or covered with snow. In the humid tropics forestation increases evapotranspiration year round and thus decreases temperature (*high confidence*). In tropical areas with a strong seasonality of rainfall, forestation will also increase evapotranspiration year round unless the soil becomes too dry. In all regions there is *medium confidence* that the diurnal temperature range decreases with increasing forest cover, with potentially reduced extreme values of temperature

Although there is not enough literature yet that rigorously compares both biophysical and biogeochemical effects of realistic scenarios of forestation, there is *high confidence* that, at the local scale (that is where the forest change occurs) biophysical effects on surface temperature are far more important than the effects resulting from the changes in emitted CO₂.

What is lacking in the literature as of today is an estimate of the impacts natural disturbances in forests will have on local climates and on the build-up of atmospheric CO₂. (O'Halloran et al. 2012) for example illustrated with many examples that changes in albedo following disturbances can result in radiative forcing changes opposite to and as large as the ones resulting from the associated changes in the net release of CO₂ by land. The resulting climate effects depend on the duration of the perturbation and of the following recovery of vegetation.

2.5.2.2 Impacts of changes in land management

There have been little changes in net cropland area over the past 50 years (at the global scale) compared to continuous changes in land management (Erb et al. 2017). Similarly, in Europe change in forest management was a very significant anthropogenic land change. Management affects water, energy and GHG fluxes exchanged between the land and the atmosphere, and thus temperature and rainfall, sometimes to the same extent as changes in land cover do as discussed in (Luyssaert et al. 2014b).

The effects of irrigation, which is a practice that has been substantially studied, and one attempt to manage solar radiation via increases in cropland albedo (geoengineering the land) is assessed, along with discussion of recent findings on the effects of forest management on local climate, although there is not enough literature yet on this topic to carry out a real assessment. The effects of urbanisation on climate are assessed in a specific cross-chapter box within this chapter (Cross-Chapter Box 4 : Climate change and urbanisation, in this chapter).

There are a number of other practices that exist, some of them being reported in Section 2.6 and chapter 6 whose importance for climate mitigation has been examined. There is however not enough literature available for assessing their biophysical effect on climate. Few papers are generally found per agricultural practice, e.g., (Jeong et al. 2014b) for double cropping, (Bagley et al. 2017) for the timing of the growing season, and (Erb et al. 2017) for a review of ten management practices.

Similarly there are very few studies that have examined how choosing species varieties and harvesting strategies in forest management impacts climate through biophysical effects, and how those effects compare to the consequences of the chosen strategies on the net CO₂ sink of the managed forest. The modelling

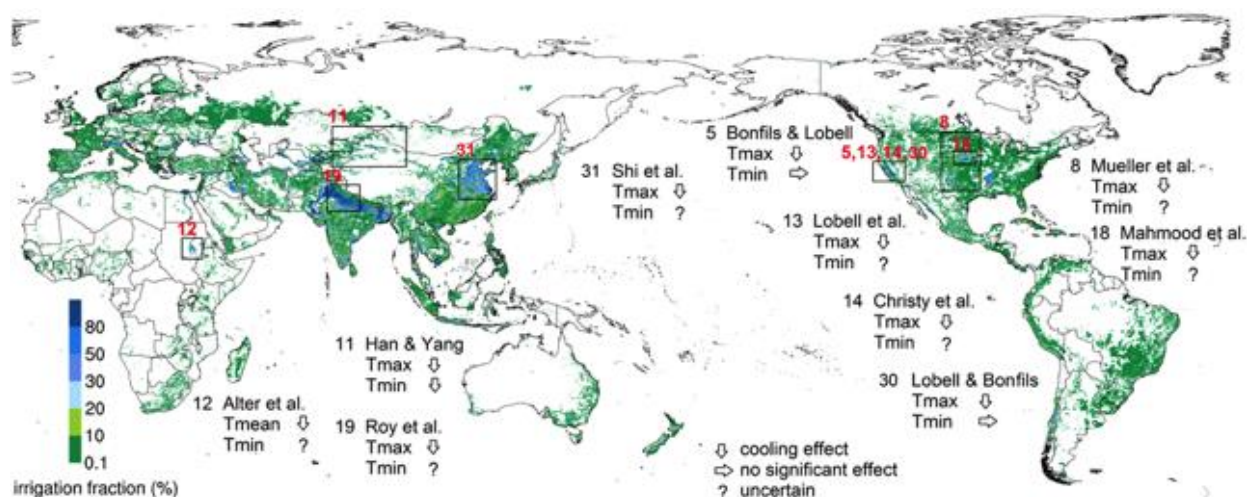
1 studies highlight the existence of competing effects between e.g. the capacity of certain species to store more
2 carbon than others (thus inducing cooling) while at the same time reducing the total evapotranspiration loss
3 and absorbing more solar radiation via lower albedo (thus inducing warming) (Naudts et al. 2016a; Luysaert
4 et al. 2018).

6 2.5.2.2.1 Irrigation

7 There is substantial literature on the effects of irrigation on local, regional and global climate as this is a
8 major land management. There is *very high confidence* that irrigation increases total evapotranspiration,
9 increases the total amount of water vapour in the atmosphere, and decreases mean surface daytime
10 temperature within the irrigated area and during the time of irrigation (Bonfils and Lobell 2007; Alter et al.
11 2015; Chen and Jeong 2018; Christy et al. 2006; Im and Eltahir 2014; Im et al. 2014; Mueller et al. 2015).
12 Decreases in maximum daytime temperature can locally be as large as -3°C to -8°C (Cook et al. 2015; Han
13 and Yang 2013; Huber et al. 2014; Alter et al. 2015; Im et al. 2014). Estimates of the contribution of
14 irrigation to past historical trends in ambient air temperature vary between -0.07°C and $-0.014^{\circ}\text{C}/\text{decade}$ in
15 Northern China (Han and Yang 2013; Chen and Jeong 2018) while being quite larger in California (-0.14°C
16 to $-0.25^{\circ}\text{C}/\text{decade}$; (Bonfils and Lobell 2007)). Surface cooling results from increased energy being taken up
17 from the land via larger evapotranspiration rates. In addition, there is growing evidence from modelling
18 studies that such cooling can locally mitigate the effect of heatwaves (Thiery et al. 2017; Mueller et al.
19 2015).

21 There is *no agreement* on changes in nighttime temperatures as discussed in (Chen and Jeong 2018) who
22 summarised the findings from observations in many regions of the World (India, China, North America and
23 eastern Africa; Figure 2.18). Where nighttime warming is found (Chen and Jeong 2018; Christy et al. 2006),
24 two explanations are put forward: the first is an increase in incoming long-wave radiation in response to
25 increased atmospheric water vapour content (greenhouse effect); the second is an increased storage of heat in
26 the soil during daytime, because of the larger heat capacity of a moister soil, heat that is then released to the
27 atmosphere at night.

29 There is *robust evidence* from modelling studies that implementing irrigation enhances rainfall although
30 there is *very low confidence* on where this increase occurs. When irrigation occurs in Sahelian Africa, during
31 the monsoon period, rainfall is decreased over the irrigated areas (*high agreement*) and increases south-west
32 if the crops are located in western Africa (Alter et al. 2015) and east / north-east when crops are located
33 further East in Sudan (Im and Eltahir 2014; Im et al. 2014) The cooler irrigated surfaces in the Sahel,
34 because of their greater evapotranspiration, inhibits convection and creates an anomalous descending motion
35 over crops that suppresses rainfall but influences the circulation of monsoon winds. Irrigation in India occurs
36 prior to the start of the monsoon season and the resulting land cooling decreases the land-sea temperature
37 contrast. This can delay the onset of the Indian monsoon and decrease its intensity (Niyogi et al. 2010;
38 Guimberteau et al. 2012). Results from (De Vrese et al. 2016a) modelling study suggest that part of the
39 excess rainfall triggered by Indian irrigation falls westward, in the horn of Africa. The theory behind those
40 local and downwind changes in rainfall support the findings from the models but we do not yet have
41 sufficient literature to robustly assess the magnitude and exact location of the expected changes driven by
42 irrigation.



1
2 **Figure 2.18** Global map of areas equipped for irrigation (colours), expressed as a percentage of total area, or
3 irrigation fraction (Siebert et al. 2013). Numbered boxes show regions where irrigation causes cooling (down
4 arrow) of surface mean (Tmean), maximum (Tmax) or minimum (Tmin) temperature, or else no significant
5 effect (right arrow) or where the effect is uncertain (question mark), based on observational studies as
6 reviewed in (Chen and Jeong 2018). Tmax refers to the warmest daily temperature while Tmin to the coldest
7 one which generally occurs at night. References are (Alter et al. 2015; Han and Yang 2013; Roy et al. 2007;
8 Shi et al. 2013; Bonfils and Lobell 2007; Lobell et al. 2008; Lobell and Bonfils 2008; Christy et al. 2006;
9 Mahmood et al. 2006; Mueller et al. 2015)

11 2.5.2.2.2 Cropland albedo

12 Various methods have been proposed to increase surface albedo in cropland and thus reduce locally surface
13 temperature (*high confidence*): choose ‘brighter’ crop varieties (Ridgwell et al. 2009; Crook et al. 2015;
14 Hirsch et al. 2017; Singarayer et al. 2009; Singarayer and Davies-Barnard 2012), abandon tillage (Lobell et
15 al. 2006; Davin et al. 2014), include cover crops into the rotation in areas where soils are darker than
16 vegetation (Carrer et al. 2018; Kaye and Quemada 2017) or use greenhouses (as in (Campra et al. 2008), see
17 (Seneviratne et al. 2018) for a review).

18
19 Whatever the solution chosen, the induced reduction in absorbed solar radiation cools the land, more
20 specifically during the hottest summer days ((Davin et al. 2014; Wilhelm et al. 2015); *low confidence*)
21 (Figure 2.19). Changes in temperature are essentially local and seasonal (limited to crop growth season) or
22 sub-seasonal (when resulting from inclusion of cover crop or tillage suppression). Such management action
23 on incoming solar radiation thus holds the potential to counteract warming in cultivated areas during crop
24 growing season.

25
26 Introducing cover crops into a rotation can also have a warming effect in areas where vegetation has a darker
27 albedo than soil, or in winter during snow periods if the cover crops or their residues are tall enough to
28 overtop the snow cover (Kaye and Quemada 2017; Lombardozzi et al. 2018). In addition evapotranspiration
29 greater than that of bare soil during this transitional period reduces soil temperature (Ceschia et al. 2017).
30 Such management strategy can have another substantial mitigation effect as it allows to store carbon in the
31 soil and to reduce both direct and indirect N₂O emissions (Basche et al. 2014; Kaye and Quemada 2017), in
32 particular if fertilisation of the subsequent crop is reduced (Constantin et al. 2010, 2011). The use of cover
33 crops thus improves substantially the GHG budget of croplands (Kaye and Quemada 2017; Tribouillois et al.
34 2018). More discussion on the role of the management practices for mitigation can be found in section 2.6
35 and chapter 6.

36
37 Only a handful of modelling studies have looked at effects other than changes in atmospheric temperature in
38 response to increased cropland albedo. (Seneviratne et al. 2018) have found significant changes in rainfall
39 following an idealised increase in cropland albedo, especially within the Asian monsoon regions. The
40 benefits of cooler temperature on production, resulting from increased albedo, is cancelled by decreases in
41 rainfall that are harmful for crop productivity. The rarity of a concomitant evaluation of albedo management
42 impact on crop productivity prevents us from providing a robust assessment of this practice in terms of both
43 climate mitigation and food security.

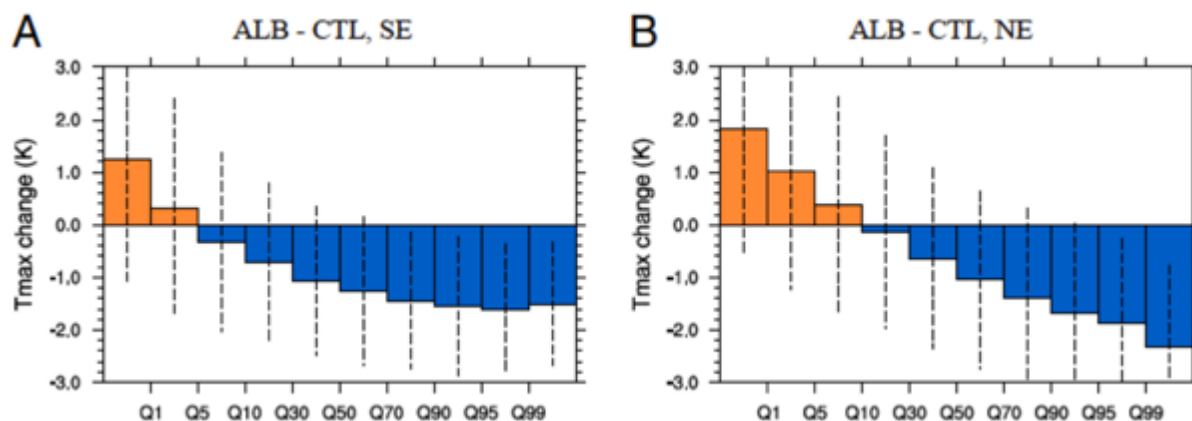
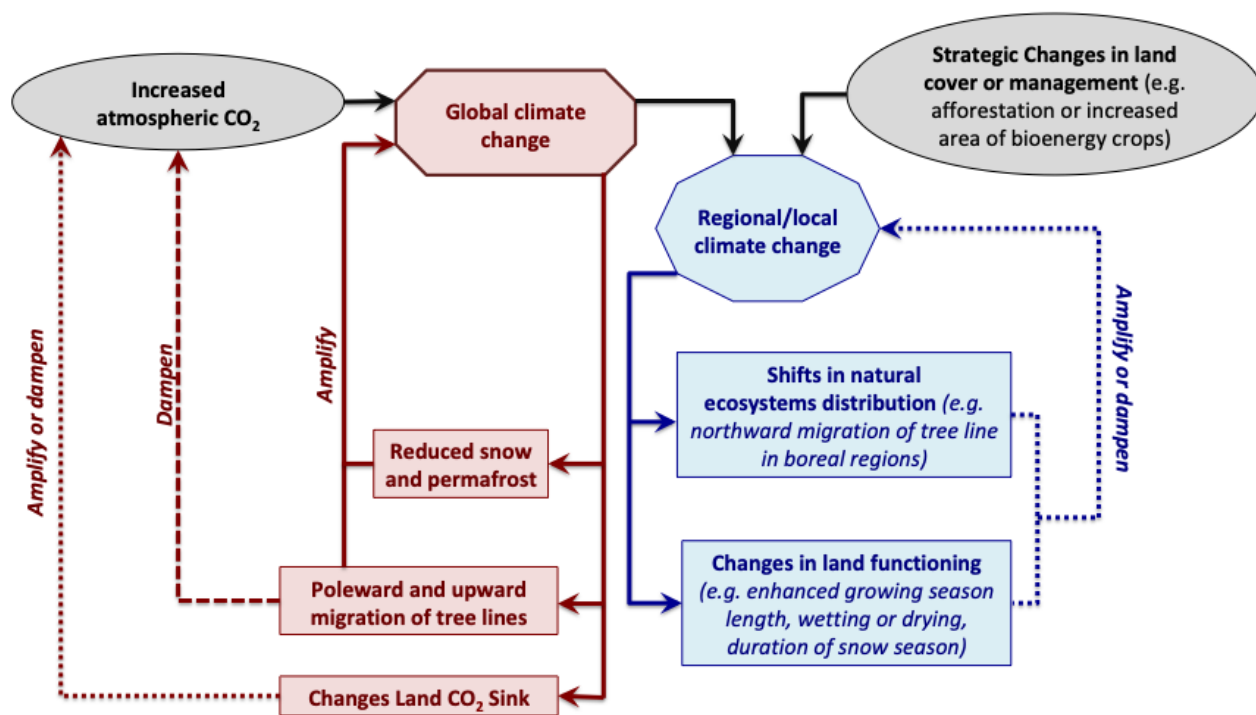


Figure 2.19 Change in summer (July-August) daily maximum temperature ($^{\circ}\text{C}$) resulting from increased surface albedo in unploughed versus ploughed land, in (A) Southern and (B) Northern Europe, during the period 1986–2009. Changes are simulated for different quantiles of the daily maximum temperature distribution, where Q1 represents the coolest 1% and Q99 the warmest 1% of summer days. Only grid cells with more than 60% of their area in cropland are included. The dashed bars represent the standard deviation calculated across all days and grid points. SE refers to southern Europe (below 45°N) and NE to northern Europe (above 45°N)

2.5.3 Amplifying / dampening climate changes via land responses

Section 2.1 and Box 2.1: illustrates the various mechanisms through which land can affect the atmosphere and thereby climate and weather. Section 2.2 illustrates the many impacts climate changes have on the functioning of land ecosystems. Section 2.3 discusses the effects future climatic conditions on the capacity of the land to absorb anthropogenic CO_2 , which then controls the sign of the feedback to the initial global warming. Sections 2.5.1 and 2.5.2 show effects of changes in anthropogenic land cover or land management on climate variables or processes. Land has thus the potential to dampen or amplify the GHG-induced global climate warming or can be used as a tool to mitigate regional climatic consequences of global warming such as extreme weather events, in addition to increasing the capacity of land to absorb CO_2 (Figure 2.20).

Land-to-climate feedbacks are difficult to assess with global or regional climate models as both types of models generally omit a large number of processes. Among these are 1) the response of vegetation to climate change in terms of growth, productivity, and geographical distribution, 2) the dynamics of major disturbances such as fires, 3) the nutrients dynamics, and 4) the dynamics and effects of short-lived chemical tracers such as biogenic volatile organic compounds (Section 2.4). Therefore, only those processes that are fully accounted for in climate models are considered here.

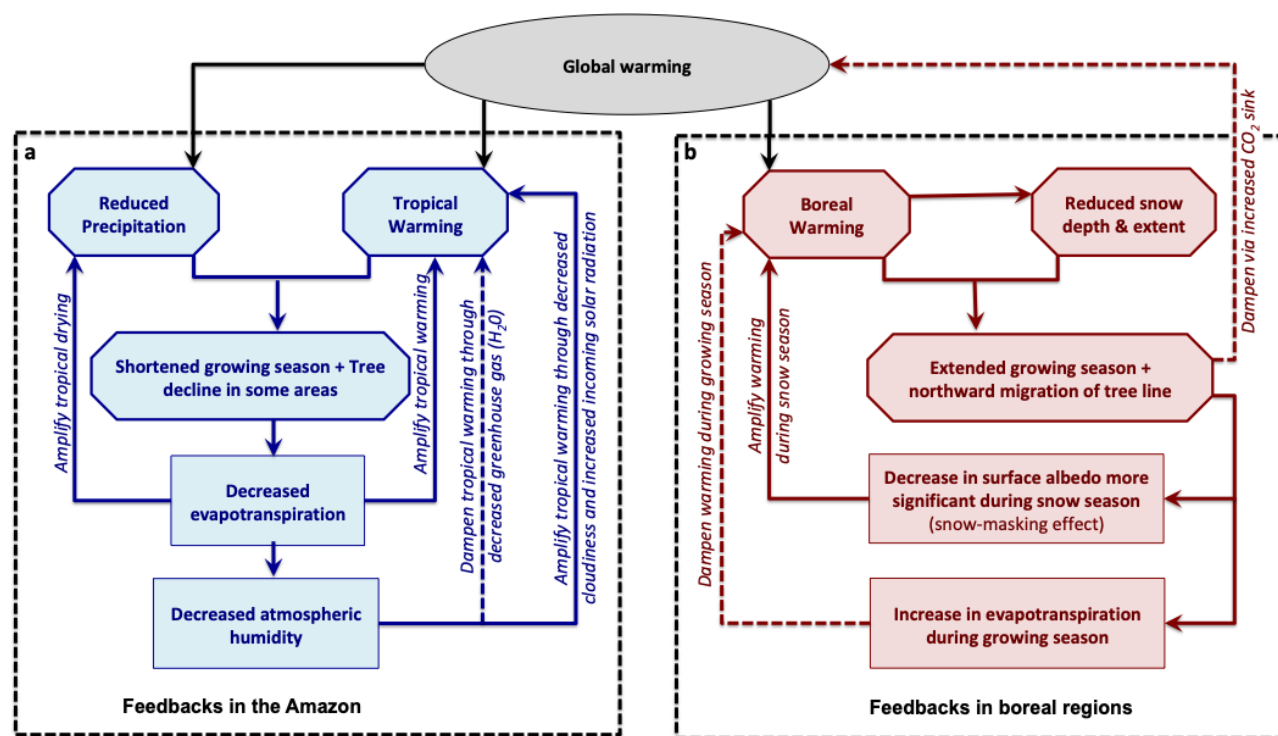


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Figure 2.20 Schematics of the various ways land has been shown, in the literature, to either amplify or dampen the initial GHG-induced climatic change, at the global scale (left panel and red boxes and arrows) or at the regional/local levels (right panel and blue boxes and arrows). Grey arrows and boxes refer to what we consider herein as imposed changes, that is the initial atmospheric GHG content as well as anthropogenic land cover change and land management. Dampening feedbacks are represented with dashed lines, amplifying ones with solid lines and the feedbacks for which the direction may be variable are represented using dotted lines. The feedbacks initiated by changes in snow and permafrost areas in boreal regions is discussed in Section 2.5.3.2, the ones initiated by changes in ecosystem distribution are discussed in Sections 2.5.3.1, 2.5.1 and 2.5.2, and the feedbacks related to changes in the land functioning are discussed in Sections 2.5.3.3, 2.5.1, as well as 2.3 and 2.5 (for changes in net CO₂ fluxes). References supporting this figure can be found in each of those sections.

2.5.3.1 Effects of changes in land cover and productivity resulting from global warming

In boreal regions, the combined northward migration of the treeline and increased growing season length in response to increased temperatures in those regions (see Section 2.2) will have positive feedbacks both on global and regional annual warming (*high confidence*; Garnaud and Sushama 2015; Jeong et al. 2014a; O’ishi and Abe-Ouchi 2009; Port et al. 2012; Strengers et al. 2010). The warming resulting from the decreased surface albedo remains the dominant signal in all modelling studies at the annual time scale and during the snow season, while cooling is obtained during the growing season (see Section 2.5.2.1; Figure 2.21 right panel).



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Figure 2.21 Schematic illustration of the processes through which the effects of global warming in the a) Amazon (left panel, blue boxes and arrows) and b) boreal regions (right panel, red boxes and arrows) feedback on the regional climate change. In boreal regions the sign of the feedbacks depends on the season, although annually global warming is further enhanced in those regions. Dashed lines illustrate negative feedbacks while solid lines indicate positive feedbacks. References supporting this figure can be found in the text.

In the tropics climate change will cause both greening and browning (see Section 2.2). Where global warming provokes decrease in rainfall, the induced decrease in biomass production leads to increased local warming (Port et al. 2012; Wu et al. 2016; Yu et al. 2016; *high confidence*). The reverse is true where warming generates increases in rainfall and thus greening. As an example, (Port et al. 2012) simulated decreases in tree cover and shortened growing season in the Amazon, despite the CO₂ fertilisation effects, in response to both future tropical warming and reduced precipitation (Figure 2.21, left panel). This browning of the land decreases both evapotranspiration and atmospheric humidity. The warming driven by the drop in evapotranspiration is enhanced via decreases in cloudiness that increases incoming solar radiation, and is dampened by reduced water vapour greenhouse radiation.

There is *very low confidence* on how feedbacks affect rainfall in the tropics where vegetation changes may occur, as the sign of the change in precipitation depends on where the greening occurs and on the season (as discussed in Section 2.5.2). There is however *high confidence* that increased vegetation growth in the southern Sahel increases African monsoon rains (Yu et al. 2016; Port et al. 2012; Wu et al. 2016). Confidence on the direction of such feedbacks is also based on a significant number of paleoclimate studies that analysed how vegetation dynamics helped maintain a northward position of the African monsoon during the Holocene time period (9 to 6 kyr BP) (de Noblet-Ducoudré et al. 2000; Rachmayani et al. 2015).

2.5.3.2 Feedbacks to climate from high-latitude land-surface changes

In high latitudes, snow albedo and permafrost carbon feedbacks are the most well-known and most important surface-related climate feedbacks because of their large-scale impacts.

In response to ongoing and projected decrease in seasonal snow cover (Derksen and Brown 2012; Brutel-Vuilmet et al. 2013) warming is and will continue to be enhanced in boreal regions (*high confidence*; Brutel-Vuilmet et al. 2013; Perket et al. 2014; Thackeray and Fletcher 2015; Mudryk et al. 2017). One reason for this is the large reflectivity (albedo) the snow exerts on shortwave radiative forcing: the all-sky global land snow shortwave radiative effect is evaluated to be around $-2.5 \pm 0.5 \text{ W m}^{-2}$ (Flanner et al. 2011; Singh et al. 2015). In the Southern Hemisphere, perennial snow on the Antarctic is the dominant contribution, while in the Northern Hemisphere, this is essentially attributable to seasonal snow with a smaller contribution from

1 snow on glaciated areas. Another reason is the sensitivity of snow cover to temperature: (Mudryk et al. 2017)
2 recently showed that in the high latitudes, climate models tend to correctly represent this sensitivity, while in
3 mid-latitude and alpine regions, the simulated snow cover sensitivity to temperature variations tends to be
4 biased low. In total, the global snow albedo feedback is about $0.1 \text{ W m}^{-2} \text{ K}^{-1}$, which amounts to about 7% of
5 the strength of the globally dominant water vapour feedback (e.g., (Thackeray and Fletcher 2015)). While
6 climate models do represent this feedback, a persistent spread in the modelled feedback strength has been
7 noticed (Qu and Hall 2014) and, on average, the simulated snow albedo feedback strength tends to be
8 somewhat weaker than in reality (Flanner et al. 2011; Thackeray and Fletcher 2015) (*medium confidence*).
9 Various reasons for the spread and biases of the simulated snow albedo feedback have been identified,
10 notably inadequate representations of vegetation masking of snow in forested areas (Loranty et al. 2014;
11 Wang et al. 2016c; Thackeray and Fletcher 2015).

12
13 The second most important potential feedback from land to climate relates to permafrost decay. There is *high*
14 *confidence* that, following permafrost decay from a warming climate, the resulting emissions of carbon
15 dioxide and/or methane (caused by the decomposition of organic matter in previously frozen soil) will
16 produce additional GHG-induced warming. There is however substantial uncertainty on the magnitude of
17 this feedback, although recent years have seen large progress in its quantification. Lack of agreement results
18 from several critical factors that carry large uncertainties. The most important are a) the size of the
19 permafrost carbon pool, b) its decomposability, c) the magnitude, timing and pathway of future high-latitude
20 climate change and d) the correct identification and model representation of the processes at play (Schuur et
21 al. 2015b). The most recent comprehensive estimates establish a total soil organic carbon storage in
22 permafrost of about $1500 \pm 200 \text{ Pg C}$ (Hugelius et al. 2014, 2013; Olefeldt et al. 2016), which is about 300
23 Pg C lower than previous estimates (*low confidence*). Important progress has been made in recent years at
24 incorporating permafrost-related processes in complex Earth System Models (e.g., (McGuire et al. 2018)),
25 but representations of some critical processes such as thermokarst formation are still in their infancy (Schuur
26 et al. 2015b). Recent model-based estimates of future permafrost carbon release (Koven et al. 2015; McGuire
27 et al. 2018) have converged on an important insight. Their results suggest that substantial net carbon release
28 of the coupled vegetation-permafrost system will probably not occur before about 2100 because carbon
29 uptake by increased vegetation growth will initially compensate for GHG releases from permafrost (*limited*
30 *evidence, high agreement*).

31 **2.5.3.3 Feedbacks related to changes in soil moisture resulting from global warming**

32 There is *medium evidence* but *high agreement* that soil moisture conditions influence the frequency and
33 magnitude of extremes such as drought and heat waves. Observational evidence indicates that dry soil
34 moisture conditions favour heat-waves, in particular in regions where evapotranspiration is limited by
35 moisture availability (Mueller and Seneviratne 2012; Quesada et al. 2012; Miralles et al. 2018; Geirinhas et
36 al. 2018; Miralles et al. 2014; Chiang et al. 2018; Dong and Crow 2019; Hirschi et al. 2014).

37
38
39 In future climate projections, soil moisture plays an important role in the projected amplification of extreme
40 heat-waves and drought in many regions of the world (*medium confidence*; (Seneviratne et al. 2013; Vogel et
41 al. 2017; Donat et al. 2018; Miralles et al. 2018)). In addition, the areas where soil moisture affects heat
42 extremes will not be located exactly where they are today. Changes in rainfall, temperature and thus
43 evapotranspiration will induce changes in soil moisture and therefore of where temperature and latent heat
44 flux will be negatively coupled (Seneviratne et al. 2006; Fischer et al. 2012). Quantitative estimates of the
45 actual role of soil moisture feedbacks are however very uncertain due to the *low confidence* in projected soil
46 moisture changes (IPCC 2013a), to weaknesses in the representation of soil moisture-atmosphere
47 interactions in climate models (Sippel et al. 2017; Ukkola et al. 2018; Donat et al. 2018; Miralles et al. 2018)
48 and to methodological uncertainties associated with the soil moisture prescription framework commonly
49 used to disentangle the effect of soil moisture on changes in temperature extremes (Hauser et al. 2017).

50
51 Where soil moisture is predicted to decrease in response to climate change in the subtropics and temperate
52 latitudes, this drying could be enhanced by the existence of soil moisture feedbacks (*low confidence* (Berg et
53 al. 2016)). The initial decrease in precipitation and increase in potential evapotranspiration and latent heat
54 flux, in response to global climate change, leads to decreased soil moisture at those latitudes and can
55 potentially amplify both. Such a feature is consistent with evidence that in a warmer climate land and
56 atmosphere will be more strongly coupled via both the water and the energy cycles (Dirmeyer et al. 2014;

1 Guo et al. 2006). This increased sensitivity of atmospheric response to land perturbations implies that
 2 changes in land uses and cover are expected, in the future, to have more impact on climate in the future than
 3 they do today.

4
 5 Beyond temperature, it has been suggested that soil moisture feedbacks influence precipitation occurrence
 6 and intensity. But the importance and even the sign of this feedback is still largely uncertain and debated
 7 (Tuttle and Salvucci 2016; Yang et al. 2018; Froidevaux et al. 2014; Guillod et al. 2015).

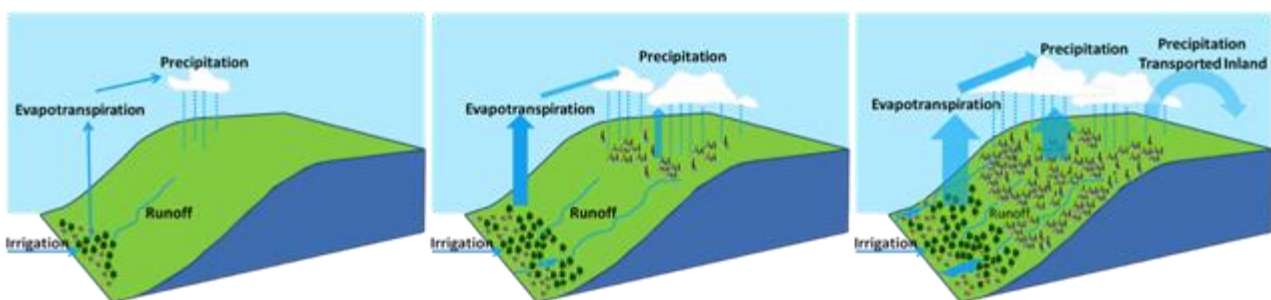
8 9 **2.5.4 Non-local and downwind effects resulting from changes in land cover**

10 Changes in land cover or land management do not just have local consequences but also affect adjacent or
 11 more remote areas. Those non-local impacts may occur in three different ways.

12
 13
 14 (1) Any action on land that affects photosynthesis and respiration has an impact on the atmospheric CO₂
 15 content as this GHG is well mixed in the atmosphere. This change in turn affects the downwelling long-wave
 16 radiation everywhere on the planet and contributes to global climate change. This is more thoroughly
 17 discussed in Section 2.6 where various land-based mitigation solutions are examined. Local land use changes
 18 thus have the potential to affect global climate via changes in atmospheric CO₂.

19
 20 (2) Any change in land cover or land management may impact local surface air temperature and moisture
 21 and thus sea-level pressure. Thermal, moisture and surface pressure gradients between the area of change
 22 and neighbouring areas are then modified and affect the amount of heat, water vapour and pollutants flowing
 23 out (downwind) of the area (e.g. Ma et al. 2013b; McLeod et al. 2017; Abiodun et al. 2012; Keys 2012).

24 Forests for example provide water vapour to the atmosphere which supports terrestrial precipitation
 25 downwind (Ellison et al. 2017a; Layton and Ellison 2016; Spracklen et al. 2012, 2018). Within a few days
 26 water vapour can travel several hundreds of kilometres before being condensed into rain and potentially
 27 being transpired again (Makarieva et al. 2009). This cascading moisture recycling (succession of
 28 evapotranspiration, water vapour transport and condensation-rainfall) has been observed in south America
 29 (Spracklen et al. 2018; Zemp et al. 2014; Staal et al. 2018; Spracklen et al. 2012). Deforestation can thus
 30 potentially decrease rainfall downwind, while combining ‘small-scale’ forestation and irrigation in the semi-
 31 arid region is susceptible to boost the precipitation-recycling mechanism with better vegetation growth
 32 downwind (Figure 2.22; (Ellison et al. 2017a; Layton and Ellison 2016)).
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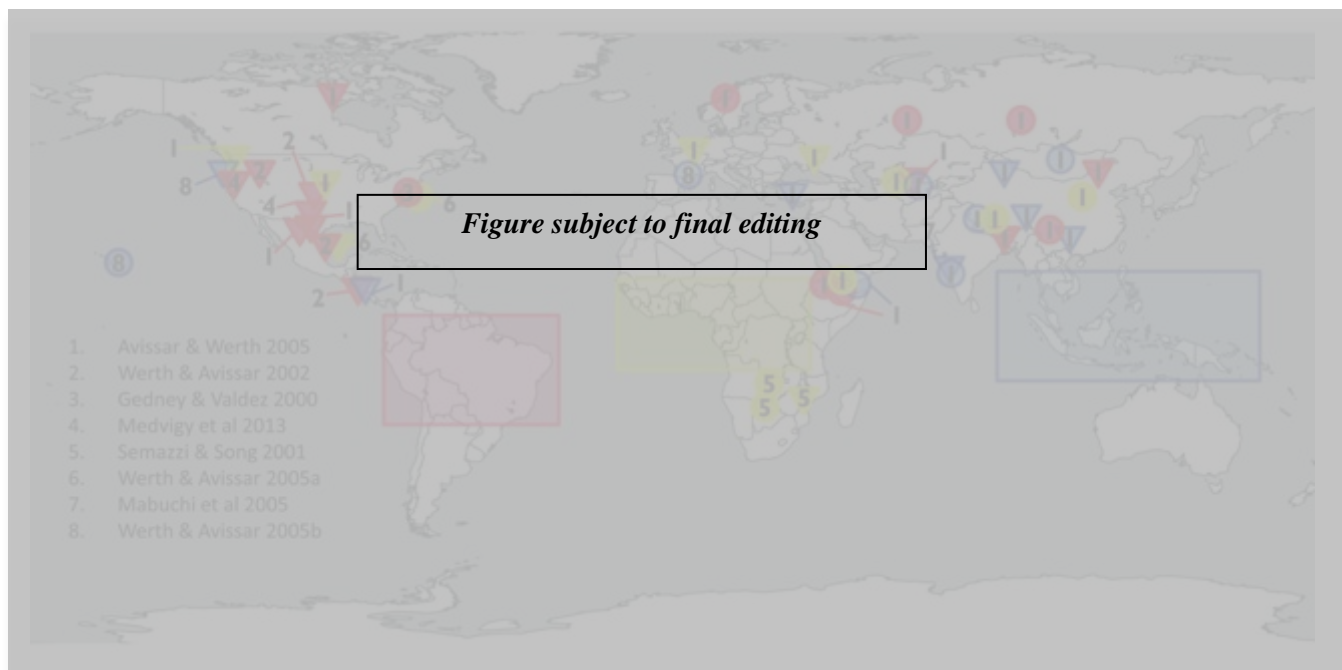


34
 35 **Figure 2.22 Schematic illustration of how combined forestation and irrigation can influence downwind**
 36 **precipitation on mountainous areas (here in Los Angeles, California area), favour vegetation growth and**
 37 **feeds back to the forested area via increased runoff (Layton and Ellison 2016). Areas of forests**
 38 **plantation and irrigation are located on the left panel, whereas consequent downwind effects and**
 39 **feedbacks are illustrated in the middle and right panels.**
 40

41 (3) Many studies using global climate models have reported that the climatic changes resulting from changes
 42 in land are not limited to the lower part of the atmosphere but can reach the upper levels via changes in large
 43 scale ascent (convection) or descent (subsidence) of air. This coupling to the upper atmosphere triggers
 44 perturbations in large-scale atmospheric transport (of heat, energy and water) and subsequent changes in
 45 temperature and rainfall in regions located quite far away from the original perturbation (Figure 2.23, Laguë
 46 and Swann 2016; Feddema et al. 2005, badger & dirmeyer 2016, Garcia 2016, Stark 2015, Devaraju 2018,
 47 Quesada et al. (2017a)).
 48

49 De Vrese et al. (2016) for example, using a global climate model, found that irrigation in India could affect

1 regions as remote as eastern Africa through changes in the atmospheric transport of water vapour. At the
 2 onset of boreal spring (February to March) evapotranspiration is already large over irrigated crops and the
 3 resulting excess moisture in the atmosphere is transported south-westward by the low-level winds. This
 4 results in increases in precipitation as large as 1mm d⁻¹ in the horn of Africa. Such finding implies that if
 5 irrigation is to decrease in India, rainfall can decrease in eastern Africa where the consequences of drought
 6 are already disastrous.
 7



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 9
 10 **Figure 2.23 Extra-tropical effects on precipitation due to deforestation in each of the three major tropical**
 11 **regions. Increasing (circles) and decreasing (triangles) precipitation result from complete deforestation of either**
 12 **Amazonia (red), Africa (yellow), or Southeast Asia (blue) as reviewed by (Lawrence and Vandecar 2015). Boxes**
 13 **indicate the area in which tropical forest was removed in each region. Numbers refer to the study from which**
 14 **the data were derived. Cited papers are the following (Avissar and Werth 2005; Gedney and Valdes 2000;**
 15 **Semazzi and Song 2001; Werth 2002; Mabuchi et al. 2005; Werth 2005)**
 16

17 Changes in sea-surface temperature have also been simulated in response to large-scale vegetation changes
 18 (Cowling et al. 2009; Davin and de Noblet-Ducoudre 2010; Wang et al. 2014b, Notaro Liu 2007).
 19 Most of those modelling studies have been carried out with land cover changes that are extremely large and
 20 often exaggerated with respect to reality. The existence of such teleconnections can thus be biased as
 21 discussed in Lorenz et al. (2016).
 22

23 In conclusion, there is *high confidence* that any action on land (for example to dampen global warming
 24 effects), wherever they occur, will not only have effects on local climate but also generate atmospheric
 25 changes in neighbouring regions, and potentially as far as few hundreds of kilometres downwind. More
 26 remote teleconnections, thousands of kilometres away from the initial perturbation, are impossible to observe
 27 and have only been reported by modelling studies using extreme land cover changes. There is *very low*
 28 *confidence* that detectable changes due to such long-range processes can occur.
 29
 30
 31

32 **Cross-Chapte Box 4: Climate Change and Urbanisation**

33
 34 Nathalie de Noblet-Ducoudré (France), Peng Cai (China), Sarah Connors (France/United Kingdom), Martin
 35 Dallimer (United Kingdom), Jason Evans (Australia), Rafiq Hamdi (Belgium), Gensuo Jia (China), Kaoru
 36 Kitajima (Japan), Christopher Lennard (South Africa), Shuaib Lwasa (Uganda), Carlos Fernando Mena
 37 (Ecuador), Soojeong Myeong (The Republic of Korea), Lennart Olsson (Sweden), Prajal Pradhan
 38 (Nepal/Germany), Lindsay Stringer (United Kingdom)

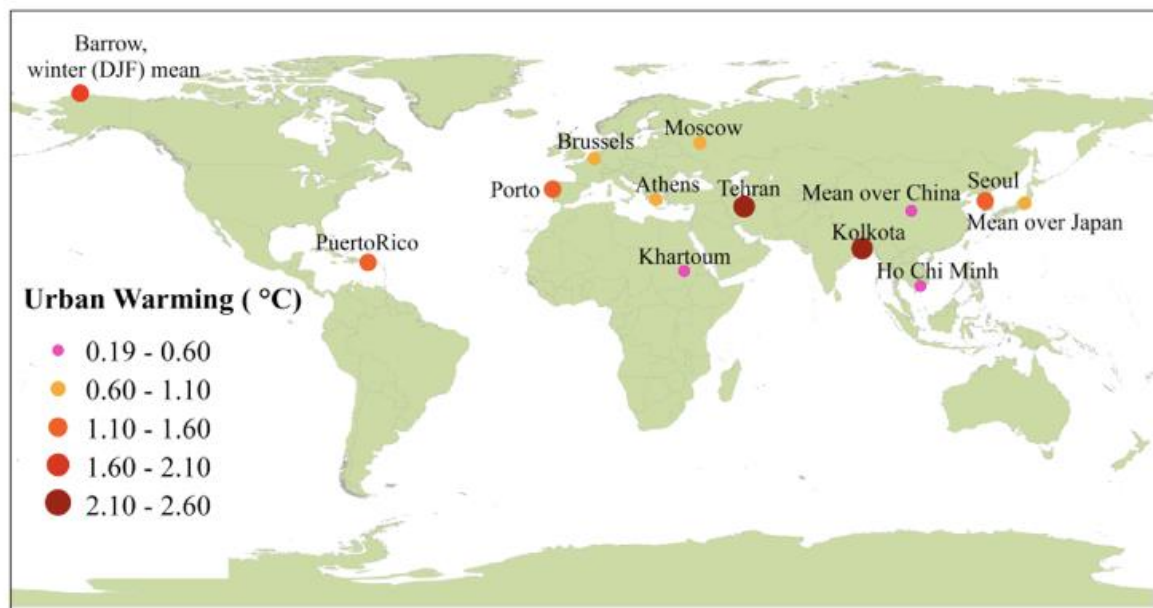
Cities extent, population, and expected growth

Despite only covering 0.4-0.9% of the global land surface (Esch et al. 2017; Zhou et al. 2015), over half the world's population live in towns and cities (United Nations 2017) generating around three-quarters of the global total carbon emissions from energy use (Creutzig et al. 2015b; Intergovernmental Panel on Climate Change 2014). Urban food consumption is a large source of these anthropogenic greenhouse gas emissions (Goldstein et al. 2017). In developed countries, per capita emissions are larger in small cities than bigger ones, while the opposite is found in developing countries (Gudipudi et al. 2019). Climate change is expected to increase the energy demand of people living in urban areas (Santamouris et al. 2015; Wenz et al. 2017).

In addition to being a driver of emissions, urbanisation contributes to forest degradation, converts neighbouring agricultural, forested, or otherwise undeveloped land to urban use, altering natural or semi-natural ecosystems both within and outside of urban areas (Du and Huang 2017). It has been identified as a major driver of land degradation as illustrated in Chapters 3, 4 and 5. Highly productive lands are experiencing the highest rate of conversion to urbanised landscapes (Nizeyimana et al. 2001; Pandey et al. 2018), affecting food security. Loss of agricultural land, and increased pollution and waste are some of key challenges arising from urbanisation and urban growth (Chen 2007). The proportion of urban population is predicted to reach ~70% by the middle of the century (United Nations 2017) with growth especially taking place in the developing world (Angel et al. 2011; Dahiya 2012). Urban sprawl is projected to consume 1.8–2.4% and 5% of the current cultivated land by 2030 and 2050 respectively (Pradhan et al. 2014; Brend'Amour et al. 2016) driven by both general population increase and migration from rural areas (Adger et al. 2015; Seto et al. 2011; Geddes et al. 2012). New city dwellers in developing countries will require land for housing to be converted from non-urban to urban land (Barbero-Sierra et al. 2013), indicating future degradation. These growing urban areas will experience direct and indirect climate change impacts, such as sea level rise and storm surges (Boettle et al. 2016; Revi et al. 2014), increasing soil salinity, and landslides from precipitation extremes. Furthermore, poorly planned urbanisation can increase people's risk to climate hazards as informal settlements and poorly built infrastructure are often the most exposed to hazards from fire, flooding, and landslides (Adger et al. 2015; Geddes et al. 2012; Revi et al. 2014). Currently, avoiding land degradation and maintaining/enhancing ecosystem services are rarely considered in planning processes (Kuang et al. 2017).

Climate change, urban heat island and threats specific to urban populations

Cities alter the local atmospheric conditions as well as those of the surrounding areas (Wang et al. 2016b; Zhong et al. 2017). There is *high confidence* that urbanisation increases mean annual surface air temperature in cities and in their surroundings, with increases ranging from 0.19°C to 2.60°C (Cross Chapter Box 4 Figure 1) (Torres-Valcárcel et al. 2015; Li et al. 2018a; Doan et al. 2016). This phenomenon is referred to as the urban heat island (UHI) effect (Oke et al. 2017; Bader et al. 2018). The magnitude and diurnal amplitude of the UHI varies from one city to another and depends on the local background climate (Wienert and Kuttler 2005; Zhao et al. 2014; Ward et al. 2016). There is nevertheless *high confidence* that urbanisation affects night time temperatures more substantially than daytime ones (Argüeso et al. 2014; Alghamdi and Moore 2015; Alizadeh-Choobari et al. 2016; Fujibe 2009; Hausfather et al. 2013; Liao et al. 2017; Sachindra et al. 2016; Camilloni and Barrucand 2012; Wang et al. 2017a; Hamdi 2010; Arsiso et al. 2018; Elagib 2011; Lokoshchenko 2017; Robaa 2013). In addition there is *high confidence* that the UHI effect makes heatwaves more intense in cities by 1.22°C to 4°C, particularly at night (Li and Bou-Zeid 2013; Li et al. 2017b; Hamdi et al. 2016; Founda and Santamouris 2017; Wang et al. 2017a). As there is a well-established relationship between extremely high temperatures and morbidity, mortality (Watts et al. 2015) and labour productivity (Costa et al. 2016), expected increase in extreme heat events with future climate change will worsen the conditions in cities.



1
2 **Cross Chapter Box 4, Figure 1: Change in annual mean surface air temperature resulting from urbanisation**
3 **(°C). Colour and size of the circles refer to the magnitude of the change. This map has been compiled using the**
4 **following studies: (Kim et al. 2016; Sun et al. 2016; Chen et al. 2016a; Founda et al. 2015; Rafael et al. 2017;**
5 **Hinkel and Nelson 2007; Chrysanthou et al. 2014; Dou et al. 2014; Zhou et al. 2016, 2017; Polydoros et al. 2018;**
6 **Li et al. 2018a; Bader et al. 2018; Alizadeh-Choobari et al. 2016; Fujibe 2009; Lokoshchenko 2017; Torres-**
7 **Valcárcel et al. 2015; Doan et al. 2016; Elagib 2011; Liao et al. 2017).**
8

9 Individual city case studies show that precipitation mean and extremes are increased over and downwind of
10 urban areas, especially in the afternoon and early evening when convective rise of the atmosphere is the
11 strongest (*medium confidence*). The case studies covered: different inland and coastal US cities (M. et al.
12 2014; McLeod et al. 2017; Ganeshan and Murtugudde 2015); Dutch coastal cities (Daniels et al. 2016);
13 Hamburg (Schlünzen et al. 2010); Shanghai (Liang and Ding 2017); Beijing (Dou et al. 2014); and Jakarta
14 and Kuala Lumpur (Lorenz et al. 2016). Increased aerosol concentrations however can interrupt the
15 precipitation formation process and thereby reduce heavy rainfall (Daniels et al. 2016; Zhong et al. 2017).
16 Urban areas also experience altered water cycle in other aspects, the evaporative demand for plants in cities
17 are increased by as much as 10% (Zipper et al. 2017) while high proportion of paving in cities mean that
18 surface runoff of water is high (Hamdi et al. 2011; Pataki et al. 2011). In addition, water retention is lower in
19 degraded, sealed soils beneath urban surfaces compared to intact soils. Increased surface water runoff,
20 especially when and where the rainfall intensity is likely to intensify (IPCC 2013b), leads to a greater
21 likelihood of flooding in urban areas without implementation of adaptation measures (Shade and Kremer
22 2019; Wang et al. 2013; Environmental Protection Agency 2015).
23

24 Urbanisation alters the stock size of soil organic carbon (SOC) and its stability. The conversion of vegetated
25 land to urban land results in a loss of carbon stored in plants, while stresses associated with the urban
26 environment (e.g., heat, limited water availability and pollution) reduce plant growth and survival in cities
27 (Xu et al. 2016b). Overall, carbon densities or stocks decrease from natural land areas to the urban core
28 along the rural-urban gradient (Tao et al. 2015; Zhang et al. 2015). For example the Seoul Forest Park, an
29 urban park, shows a tenfold difference in SOC stocks across its land cover types (Bae and Ryu 2015). In
30 Changchun in Northeast China, however, SOC density is higher in recreational forests within urban areas
31 compared to a production forest (Zhang et al. 2015).
32

33 Urban air pollution as an environmental risk increases with climate change. Increased air temperatures can
34 lead to reduced air quality by enhancing the formation of photochemical oxidants and increasing the
35 concentration of air pollutants such as ozone, with corresponding threats to human health (Sharma et al.
36 2013). The occurrence of bronchial asthma and allergic respiratory diseases is increasing worldwide, and
37 urban residents are experiencing poor air quality conditions more frequently than rural residents (D'Amato et
38 al. 2010). Excess morbidity and mortality related to extremely poor air quality are found in many cities

1 worldwide (Harlan and Ruddell 2011). Some emissions that lead to reduced air quality are also contributors
2 to climate change (Shindell et al. 2018; de Coninck et al. 2018).
3

4 **Urban response options for climate change, desertification, land degradation and food security**

5 Urban green infrastructure (UGI; see glossary) has been proposed as a solution to mitigate climate change
6 directly through carbon sequestration (Davies et al. 2011; Edmondson et al. 2014). However, compared to
7 overall carbon emissions from cities, its mitigation effects are likely to be small (*medium confidence*). UGI
8 nevertheless has an important role in adapting cities to climate change (Demuzere et al. 2014; Sussams et al.
9 2015; Martin et al. 2016; Gill et al. 2007; Revi et al. 2014). Adaptation through UGIs is achieved through,
10 for example: (i) reduction in air temperature (Cavan et al. 2014; Di Leo et al. 2016; Feyisa et al. 2014; Zölch
11 et al. 2016; Li et al. 2019) which can help improve human health and comfort (e.g. (Brown and Nicholls
12 2015; Klemm et al. 2015)); (ii) reduction in the energy demands of buildings through the use of green roofs
13 and walls (e.g.(Coma et al. 2017)); (iii) reduction in surface water runoff and flood risk (Zeleňáková et al.
14 2017). Given that UGI necessarily involves the retention and management of non-sealed surfaces, co-
15 benefits for land degradation will also be apparent (Murata and Kawai 2018; Scalenghe and Marsan 2009)
16 (*limited evidence, high agreement*).
17

18 Urban agriculture is one aspect of UGI that has the potential to both meet some of the food needs of cities
19 and reduce land degradation pressures in rural areas (*low confidence*; e.g. Wilhelm and Smith 2018). Urban
20 agriculture has many forms, such as backyard gardening, allotments, plants on roof-tops or balconies, urban-
21 fringe/peri-urban agriculture, hydroponics, aquaponics, livestock grazing in open spaces and vertical farming
22 (Gerster-Bentaya 2013) (see also Section 5.6.5).
23

24 Consuming locally produced food and enhancing the efficiency of food processing and transportation can
25 minimise food losses, contribute to food security, and in some circumstances reduce GHG emissions (Brodt
26 et al. 2013; Michalský and Hooda 2015; Tobarra et al. 2018) (see also Section 5.5.2.3). Furthermore, urban
27 agriculture has the potential to counteract the separation of urban populations from food production. This
28 separation is one driver of the transition towards more homogeneous, high protein diets, which are associated
29 with increased greenhouse gas emissions (Goldstein et al. 2017; Moragues-Faus and Marceau 2018;
30 Magarini and Calori 2015). Barriers to the uptake of urban agriculture as a climate change mitigation option
31 include the need for efficient distribution systems to ensure lowered carbon emissions (Newman et al. 2012)
32 and the concern that urban agriculture may harbour pathogenic diseases, or that its products be contaminated
33 by soil or air pollution (Hamilton et al. 2014; Ercilla-Montserrat et al. 2018).
34

35 **In summary**

36 Climate change is already affecting the health and energy demand of large numbers of people living in urban
37 areas (*high confidence*; see also Section 2.2). Future changes to both climate and urbanisation will enhance
38 warming in cities and their surroundings, especially during heat waves (*high confidence*). Urban and peri-
39 urban agriculture, and more generally the implementation of urban green infrastructure, can contribute to
40 climate change mitigation (*medium confidence*) as well as to adaptation (*high confidence*), including co-
41 benefits for food security and reduced soil-water-air pollution.
42
43

2.6 Climate consequences of response options

Response options can affect climate mitigation and adaptation simultaneously, therefore this Special Report on Climate Change and Land (SRCCL) discusses land-based response options in an integrated way (Chapter 1). In this chapter we assess response options that have an effect on climate. A description of the full set of response options across the SRCCL can be found in Chapter 6, including the interplay between mitigation, adaptation, desertification, land degradation, food security and other co-benefits and trade-offs. Response options specific to desertification, degradation and food security are described in more detail in Chapters 3, 4 and 5.

Some response options lead to land use change and can compete with other land uses, including other response options, while others may free-up land that can be used for further mitigation/adaptation by reducing demand for land or products e.g. agricultural intensification, diet shifts, and reduction of waste (*high confidence*).

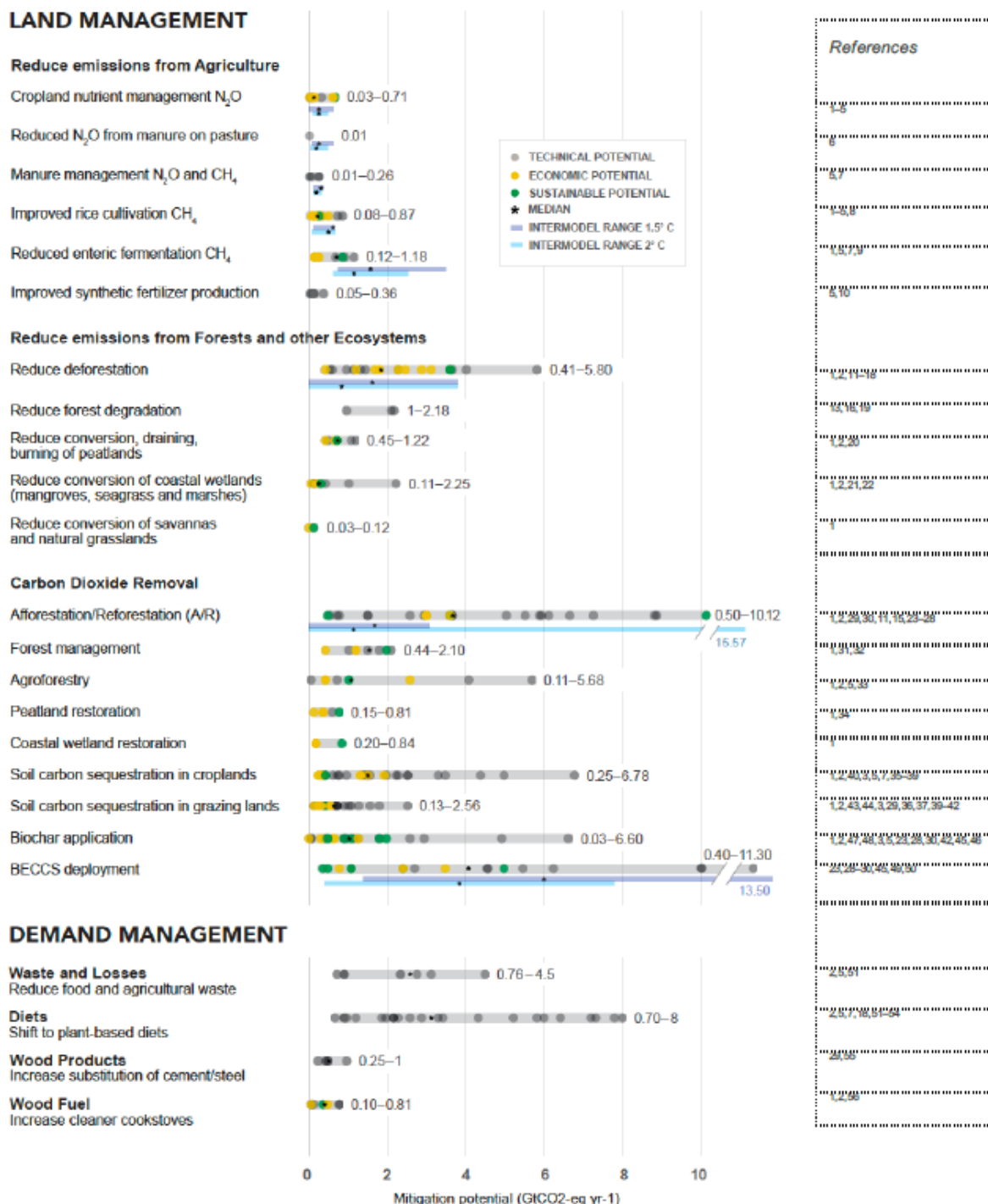
Some response options result in a net removal of GHGs from the atmosphere and storage in living or dead organic material, or in geological stores (IPCC SR15). Such options are frequently referred to in the literature as carbon dioxide removal (CDR), Greenhouse Gas Removal (GGR) or negative emissions technologies (NETs). CDR options are assessed alongside emissions reduction options. Although they have a land footprint, solar and wind farms are not assessed here as they affect greenhouse gas flux in the energy industrial sectors with minimal effect in the land sector, but the impact of solar farms on agricultural land competition is dealt with in Chapter 7.

A number of different types of scenario approach exist for estimating climate contribution of land-based response options (see Cross-Chapter Box 1: Scenarios, Chapter 1). Mitigation potentials have been estimated for single and sometimes multiple response options using stylised “bottom-up” scenarios. Response options are not mutually exclusive (e.g., management of soil carbon and cropland management). Different options interact with each other; they may have additive effects or compete with each other for land or other resources, thus these potentials cannot necessarily be added up. The interplay between different land-based mitigation options, as well as with mitigation options in other sectors (such as energy or transport), in contributing to specific mitigation pathways has been assessed using Integrated Assessment Models, see Section 2.7.2. These include interactions with wider socioeconomic conditions (see Cross-Chapter Box 1: Scenarios, Chapter 1) and other sustainability goals (see chapter 6).

2.6.1 Climate impacts of individual response options

Since AR5, there have been many new estimates of the climate impacts of single or multiple response options, summarised in Figure 2.24 and discussed in sub-sections below. Recently published syntheses of mitigation potential of land-based response options (e.g. Hawken 2017a; Smith et al. 2016b; Griscom et al. 2017a; Minx et al. 2018; Fuss et al. 2018b; Nemet et al. 2018) are also included in Figure 2.24. The wide range in mitigation estimates reflects differences in methodologies that may not be directly comparable, and estimates cannot be necessarily be added if they were calculated independently as they may be competing for land and other resources.

Some studies assess a “technical mitigation potential” - the amount possible with current technologies. Some include resource constraints (e.g., limits to yields, limits to natural forest conversion) to assess a “sustainable potential”. Some assess an “economic potential” mitigation at different carbon prices. Few include social and political constraints (e.g. behaviour change, enabling conditions, see Chapter 7), the biophysical climate effects (Section 2.5), or the impacts of future climate change (Section 2.3). Carbon stored in biomass and soils may be at risk of future climate change (see Section 2.2), natural disturbances such as wildfire (see Cross-Chapter Box 3: Fire and Climate Change, in this chapter) and future changes in land use or management changes that result in a net loss of carbon (Gren and Aklilu 2016).



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Figure 2.24 Mitigation potential of response options in 2020-2050, measured in GtCO₂-eq yr⁻¹, adapted from Roe et al. (2017). Mitigation potentials reflect the full range of low to high estimates from studies published after 2010, differentiated according to technical (possible with current technologies), economic (possible given economic constraints) and sustainable potential (technical or economic potential constrained by sustainability considerations). Medians are calculated across all potentials in categories with >4 data points. Only includes references that explicitly provide mitigation potential estimates in CO₂-eq yr⁻¹ (or similar derivative) by 2050. Not all options land management potentials are additive as some may compete for land. Estimates reflect a range of methodologies (including definitions, global warming potentials and time horizons) that may not be directly comparable or additive. Results from Integrated Assessment Models are shown to compare with single option “bottom-up” estimates, in available categories from the 2°C and 1.5°C scenarios in the SSP Database (Version 2.0). The models reflect land management changes, yet in some instances, can also reflect demand-side effects from carbon prices, so may not be defined exclusively as “supply-side”. References: 1. Griscom et al. (2017); 2. Hawken (2017); 3. Paustian et al. (2016); 4. Beach et al. (2016); 5. Dickie et al. (2014); 6. Herrero et al. (2013); 7. Herrero et al. (2016); 8. Hussain et al. (2015); 9. Hristov, et al. (2013); 10. Zhang et al. (2013); 11. Houghton &

1 Nassikas (2018); 12. Busch & Engelmann (2017); 13. Baccini et al. (2017); 14. Zarin et al. (2016); 15. Houghton,
 2 et al. (2015); 16. Federici et al. (2015); 17. Carter et al. (2015); 18. Smith et al. (2013); 19. Pearson et al. (2017);
 3 20. Hooijer et al. (2010); 21. Howard (2017); 22. Pendleton et al. (2012); 23. Fuss et al. (2018); 24. Dooley &
 4 Kartha (2018); 25. Kreidenweis et al. (2016); 26. Yan et al. (2017); 27. Sonntag et al. (2016); 28. Lenton (2014);
 5 29. McLaren (2012); 30. Lenton (2010); 31. Sasaki et al. (2016); 32. Sasaki et al. (2012); 33. Zomer et al. (2016);
 6 34. Couwenberg et al. (2010); 35. Conant et al. (2017); 36. Sanderman et al. (2017); 37. Frank et al. (2017); 38.
 7 Henderson et al. (2015); 39. Sommer & Bossio (2014); 40. Lal (2010); 41. Zomer et al. (2017); 42. Smith et al.
 8 (2016); 43. Poeplau & Don (2015); 44. Powlson et al. (2014); 45. Powell & Lenton (2012); 46. Woolf et al. (2010);
 9 47. Roberts et al. (2010); 48. Pratt & Moran (2010); 49. Turner et al. (2018); 50. Koornneef et al. (2012); 51.
 10 Bajželj et al. (2014); 52. Springmann et al. (2016); 53. Tilman & Clark (2014); 54. Hedenus et al. (2014); 55.
 11 Miner (2010); 56. Bailis et al. (2015)

15 2.6.1.1 Land management in agriculture

16 Reducing non-CO₂ emissions from agriculture through cropland nutrient management, enteric fermentation,
 17 manure management, rice cultivation and fertiliser production has a total mitigation potential of 0.30–3.38
 18 GtCO₂-eq yr⁻¹ (*medium confidence*) (combined sub-category measures in **Figure 2.24**, details below) with a
 19 further 0.25–6.78 GtCO₂-eq yr⁻¹ from soil carbon management (Section 2.6.1.3). Other literature that looks at
 20 broader categories finds mitigation potential of 1.4–2.3 GtCO₂-eq yr⁻¹ from improved cropland management
 21 (Smith et al. 2008, 2014; Pradhan et al., 2013); 1.4–1.8 GtCO₂-eq yr⁻¹ from improved grazing land
 22 management (Conant et al. 2017; Herrero et al. 2016; Smith et al. 2008, 2014) and 0.2–2.4 GtCO₂-eq yr⁻¹
 23 from improved livestock management (Smith et al. 2008, 2014; Herrero et al. 2016, FAO 2007). A detailed
 24 discussion of the mitigation potential of agricultural response options and their co-benefits are provided in
 25 Chapter 5, Section 5.5. and 5.6).

26
 27 The three main measures to reduce enteric fermentation include improved animal diets (higher quality, more
 28 digestible livestock feed), supplements and additives (reduce methane by changing the microbiology of the
 29 rumen), and animal management and breeding (improve husbandry practices and genetics) – and applying
 30 these measures can mitigate 0.12–1.18 GtCO₂-eq yr⁻¹ (*medium confidence*) (Hristov et al. 2013; Dickie et al.
 31 2014; Herrero et al. 2016; Griscom et al. 2017). However, these measures may have limitations such as need
 32 of crop-based feed (Pradhan et al. 2013) and associated ecological costs and toxicity and animal welfare
 33 issues related to food additives (Llonch et al. 2017). Measures to manage manure include anaerobic digestion
 34 for energy use, composting as a nutrient source, reducing storage time, and changing livestock diets, and
 35 have a potential of 0.01–0.26 GtCO₂-eq yr⁻¹ (Herrero et al. 2016a; Dickie et al. 2014).

36
 37 On croplands, there is a mitigation potential of 0.03–0.71 GtCO₂-eq yr⁻¹ for cropland nutrient management
 38 (fertiliser application) (*medium confidence*) (Griscom et al. 2017a; Hawken 2017; Paustian et al. 2016;
 39 Dickie et al. 2014; Beach et al. 2015). Reducing emissions from rice production through improved water
 40 management (periodic draining of flooded fields to reduce methane emissions from anaerobic
 41 decomposition), and straw residue management (apply in dry conditions instead of on flooded fields, avoid
 42 burning to reduce methane and nitrous oxide emissions) has the potential to mitigate up to 60% of emissions
 43 (Hussain et al. 2015), or 0.08–0.87 GtCO₂-eq yr⁻¹ (*medium confidence*) (Griscom et al. 2017a; Hawken
 44 2017; Paustian et al. 2016; Hussain et al. 2015; Dickie et al. 2014; Beach et al. 2015). Further, sustainable
 45 intensification through the integration of crop and livestock systems can increase productivity, decrease
 46 emission intensity and act as a climate adaptation option (see chapter 5.5.1.4).

47
 48 Agroforestry is a land management system that combines woody biomass (e.g., trees or shrubs) with crops
 49 and/or livestock). The mitigation potential from agroforestry ranges between 0.08 to 5.7 GtCO₂ yr⁻¹, (*medium*
 50 *confidence*) (Griscom et al. 2017c; Dickie et al. 2014; Zomer et al. 2016; Hawken 2017). The high estimate
 51 is from an optimum scenario combining four agroforestry solutions (silvopasture, tree intercropping,
 52 multistrata agroforestry and tropical staple trees) of Hawken (2017a). Zomer et al. (2016) reported that the
 53 trees in agroforestry landscapes had increased carbon stock by 7.33 GtCO₂ between 2000 and 2010, or 0.7
 54 GtCO₂ yr⁻¹. For more details see Chapter 5, Section 5.5.1.3.

56 2.6.1.2 Land management in forests

57 The mitigation potential for reducing and/or halting deforestation and degradation ranges from 0.4 to 5.8

1 GtCO₂ yr⁻¹ (*high confidence*) (Griscom et al. 2017a; Hawken 2017; Busch and Engelmann 2017; Baccini et
2 al. 2017; Zarin et al. 2016b; Federici et al. 2015; Carter et al. 2015; Houghton et al. 2015; Smith et al. 2013a;
3 Houghton and Nassikas 2018). The higher figure represents a complete halting of land use conversion in
4 forests and peatlands (i.e. assuming recent rates of carbon loss are saved each year). Separate estimates of
5 degradation only range from 1.0-2.18 GtCO₂ yr⁻¹. Reduced deforestation and forest degradation include
6 conservation of existing carbon pools in vegetation and soil through protection in reserves, controlling
7 disturbances such as fire and pest outbreaks, and changing management practices. Differences in estimates
8 stem from varying land cover definitions, time periods assessed, and carbon pools included (most higher
9 estimates include belowground, dead wood, litter, soil, and peat carbon). When deforestation and
10 degradation are halted, it may take many decades to fully recover the biomass initially present in native
11 ecosystems (Meli et al. 2017, See also Chapter 4.8.3).

12
13 Afforestation/Reforestation (A/R) and forest restoration can increase carbon sequestration in both vegetation
14 and soils by 0.5–10.1 GtCO₂ yr⁻¹ (*medium confidence*) (Fuss et al. 2018; Griscom et al. 2017a; Hawken
15 2017; Kreidenweis et al. 2016; Li et al. 2016; Huang et al. 2017; Sonntag et al. 2016; Lenton 2014; McLaren
16 2012; Lenton 2010; Erb et al. 2018a; Dooley and Kartha 2018; Yan et al. 2017; Houghton et al. 2015;
17 Houghton and Nassikas 2018). Afforestation is the conversion to forest of land that historically has not
18 contained forests. Reforestation is the conversion to forest of land that has previously contained forests but
19 that has been converted to some other use. Forest restoration refers to practices aimed at regaining ecological
20 integrity in a deforested or degraded forest landscape. The lower estimate represents the lowest range from
21 an earth system model (Yan et al. 2017) and of sustainable global negative emissions potential (Fuss et al.
22 2018), and the higher estimate reforests all areas where forests are the native cover type, constrained by food
23 security and biodiversity considerations (Griscom et al. 2017a). It takes time for full carbon removal to be
24 achieved as the forest grows. Removal occurs at faster rates in young to medium aged forests and declines
25 thereafter such that older forest stands have smaller carbon removals but larger stocks with net uptake of
26 carbon slowing as forests reach maturity (Yao et al. 2018; Poorter et al. 2016; Tang et al. 2014). The land
27 intensity of afforestation and reforestation has been estimated at 0.0029 km² tC⁻¹ yr⁻¹ (Smith et al. 2016a).
28 Boysen et al. (2017) estimated that to sequester about 100 GtC by 2100 would require 13 Mkm² of
29 abandoned cropland and pastures. See also Chapter 4.8.3.

30
31 Forest management has the potential to mitigate 0.4-2.1 GtCO₂-eq yr⁻¹ (*medium confidence*) (Sasaki et al.
32 2016; Griscom et al. 2017; Sasaki et al. 2012). Forest management can alter productivity, turnover rates,
33 harvest rates carbon in soil, and carbon in wood products (Erb et al. 2017; Campioli et al. 2015; Birdsey and
34 Pan 2015; Erb et al. 2016; Noormets et al. 2015; Wäldchen et al. 2013; Malhi et al. 2015; Quesada et al.
35 2018; Nabuurs et al. 2017; Bosello et al. 2009) (see also Chapter 4, Section 4.8.4). Fertilisation may enhance
36 productivity but would increase N₂O emissions. Preserving and enhancing carbon stocks in forests has
37 immediate climate benefits but the sink can saturate and is vulnerable to future climate change (Seidl et al.
38 2017). Wood can be harvested and used for bioenergy substituting for fossil fuels (with or without carbon
39 capture and storage) (Section 2.6.1.5), for long-lived products such as timber (see below), to be buried as
40 biochar (Section 2.6.1.1) or to be used in the wider bioeconomy, enabling areas of land to be used
41 continuously for mitigation. This leads to initial carbon loss and lower carbon stocks but with each harvest
42 cycle, the carbon loss (debt) can be paid back and after a parity time, result in net savings (Laganière et al.
43 2017; Bernier and Paré 2013; Mitchell et al. 2012; Haberl et al. 2012; Haberl 2013; Ter-Mikaelian et al.
44 2015; Macintosh et al. 2015). The trade-off between maximising forest C stocks and maximising substitution
45 is highly dependent on the counterfactual assumption (no-use vs. extrapolation of current management),
46 initial forest conditions and site-specific contexts such as regrowth rates and, the displacement factors and
47 efficiency of substitution, and relative differences in emissions released during extraction, transport and
48 processing of the biomass- or fossil-based resources as well as assumptions about emission associated with
49 the product or energy source that is substituted (Grassi et al. 2018b; Nabuurs et al. 2017; Pingoud et al. 2018;
50 Smyth et al. 2017a; Luysaert et al. 2018; Valade et al. 2017; York 2012; Ter-Mikaelian et al. 2014; Naudts
51 et al. 2016b; Mitchell et al. 2012; Haberl et al. 2012; Macintosh et al. 2015; Laganière et al. 2017; Haberl
52 2013). This leads to uncertainty about optimum mitigation strategies in managed forests, while high carbon
53 ecosystems such as primary forests would have large initial carbon losses and long pay-back times and thus
54 protection of stocks would be more optimal (Lemprière et al. 2013; Kurz et al. 2016; Keith et al. 2014). See
55 also 4.8.4.

1 Global mitigation potential from increasing the demand of wood products to replace construction materials
2 range from 0.25–1 GtCO₂-eq yr⁻¹ (McLaren 2012; Miner 2010) (*medium confidence*), the uncertainty is
3 determined in part by consideration of the factors described above, and is sensitive to the displacement
4 factor, or the substitution benefit in CO₂, when wood is used instead of another material, which may vary in
5 the future as other sectors reduce emissions, as well as market factors (Sathre and O'Connor 2010; Nabuurs
6 et al. 2018; Iordan et al. 2018; Braun et al. 2016; Gustavsson et al. 2017; Peñaloza et al. 2018; Soimakallio et
7 al. 2016; Grassi et al. 2018b). Using harvested carbon in long-lived products (e.g., for construction) can
8 represent a store that can sometimes be from decades to over a century while the wood can also substitute for
9 intensive building materials, avoiding emissions from the production of concrete and steel (Sathre and
10 O'Connor 2010; Smyth et al. 2017b; Nabuurs et al. 2007; Lemprière et al. 2013). The harvest of carbon and
11 storage in products affects the net carbon balance of the forest sector, with the aim of sustainable forest
12 management strategies being to optimise carbon stocks and use of harvested products to generate sustained
13 mitigation benefits (Nabuurs et al. 2007).

14
15 Biophysical effects of forest response options are variable depending on the location and scale of activity
16 (Section 2.6). Reduced deforestation or afforestation in the tropics contributes to climate mitigation through
17 both biogeochemical and biophysical effects. It also maintains rainfall recycling to some extent. In contrast,
18 in higher latitude boreal areas observational and modelling studies show that afforestation and reforestation
19 lead to local and global warming effects, particularly in snow covered regions in the winter as the albedo is
20 lower for forests than bare snow (Bathiany et al. 2010a; Dass et al. 2013; Devaraju et al. 2018b; Ganopolski
21 et al. 2001b; Snyder et al. 2004a; West et al. 2011; Arora and Montenegro 2011) (Section 2.6). Management,
22 e.g. thinning practices in forestry, could increase the albedo in regions where albedo decreases with age. The
23 length of rotation cycles in forestry affects tree height and thus roughness, and through the removal of leaf
24 mass, harvest reduces evapotranspiration (Erb et al. 2017) which could lead to increased fire susceptibility in
25 the tropics. In temperate and boreal sites, biophysical forest management effects on surface temperature were
26 shown to be of similar magnitude than changes in land cover (Luyssaert et al. 2014b). These biophysical
27 effects could be of a magnitude to overcompensate biogeochemical effects, e.g. the sink strength of
28 regrowing forests after past depletions (Luyssaert et al. 2018; Naudts et al. 2016b), but many parameters and
29 assumptions on counterfactual influence the account (Anderson et al. 2011; Li et al. 2015b; Bright et al.
30 2015).

31
32 Forest cover also affects climate through reactive gases and aerosols with *limited evidence* and *medium*
33 *agreement* that the decrease in the emissions of BVOC resulting from the historical conversion of forests to
34 cropland has, resulted in a positive radiative forcing through direct and indirect aerosol effects, a negative
35 radiative forcing through the reduction in the atmospheric lifetime of methane it has increased an decreased
36 ozone concentrations in different regions (see Section 2.4).

37 38 39 **2.6.1.3 Land management of soils**

40 The global mitigation potential for increasing soil organic matter stocks in mineral soils is estimated to be in
41 the range of 0.4–8.64 GtCO₂ yr⁻¹ (*high confidence*) though the full literature range is wider with high
42 uncertainty related to some practices (Fuss et al. 2018; Sommer and Bossio 2014; Lal 2010; Lal et al. 2004;
43 Conant et al. 2017; Dickie et al. 2014; Frank et al. 2017a; Griscom et al. 2017c; Herrero et al. 2015, 2016b;
44 McLaren 2012; Paustian et al. 2016; Poeplau and Don 2015; Powlson et al. 2014b; Smith et al. 2016d;
45 Zomer et al. 2017). Some studies have separate potentials for soil carbon sequestration in croplands (0.25-
46 6.78 GtCO₂ yr⁻¹) (Griscom et al. 2017a; Hawken 2017; Frank et al. 2017a; Paustian et al. 2016; Herrero et al.
47 2016a; Henderson et al. 2015b; Dickie et al. 2014; Conant et al. 2017; Lal 2010) and soil carbon
48 sequestration in grazing lands (0.13-2.56 GtCO₂ yr⁻¹) (Griscom et al. 2017a; Hawken 2017; Frank et al.
49 2017a; Paustian et al. 2016; Powlson et al. 2014a; McLaren 2012; Zomer et al. 2017; Smith et al. 2015;
50 Sommer and Bossio 2014; Lal 2010). The potential for soil carbon sequestration and storage varies
51 considerably, depending on prior and current land management approaches, soil type, resource availability,
52 environmental conditions, microbial composition and nutrient availability among others (Hassink and
53 Whitmore 1997; Smith and Dukes 2013; Palm et al. 2014; Lal 2013; Six et al. 2002; Feng et al. 2013). Soils
54 are a finite carbon sink and sequestration rates may decline to negligible levels over as little as a couple of
55 decades as soils reach carbon saturation (West et al. 2004; Smith and Dukes 2013). The sink is at risk of
56 reversibility, in particular due to increased soil respiration under higher temperatures (section 2.3)

1
2 Land management practices to increase carbon interact with agricultural and fire management practices (see
3 Cross-chapter box 3: Fire and Climate Change, and Chapter 5) and include improved rotations with deeper
4 rooting cultivars, addition of organic materials, and agroforestry (Lal 2011; Smith et al. 2008; Lorenz and
5 Pitman 2014; Lal 2013; Vermeulen et al. 2012a; de Rouw et al. 2010). Adoption of green manure cover
6 crops, while increasing cropping frequency or diversity, helps sequester SOC (Poepflau and Don 2015;
7 Mazzoncini et al. 2011; Luo et al. 2010). Studies of the long-term SOC sequestration potential of
8 conservation agriculture, i.e. the simultaneous adoption of minimum tillage, (cover) crop residue retention
9 and associated soil surface coverage, and crop rotations, includes results that are both positive (Powlson et
10 al. 2016; Zhang et al. 2014) and inconclusive (Cheesman et al. 2016; Palm et al. 2014; Govaerts et al. 2009).

11
12 The efficacy of reduced and zero-till practices is highly context-specific; many studies demonstrate increased
13 carbon storage (e.g. Paustian et al., 2000; Six et al., 2004; van Kessel et al., 2013), while others show the
14 opposite effect (Sisti et al. 2004; Álvaro-Fuentes et al. 2008; Christopher et al. 2009). On the other hand,
15 deep ploughing can contribute to SOC sequestration by burying soil organic matter in the subsoil where it
16 decomposes slowly (Alcántara et al. 2016). Meta-analyses (Haddaway et al. 2017; Luo et al. 2010; Meurer
17 et al. 2018) also show a mix of positive and negative responses, and the lack of robust comparisons of soils
18 on an equivalent mass basis continues to be a problem for credible estimates (Wendt and Hauser 2013;
19 Powlson et al. 2011; Powlson et al. 2014).

20
21 Soil carbon management interacts with N₂O (Paustian et al. 2016). For example, (Li et al. 2005) estimate that
22 the management strategies required to increase C sequestration (reduced tillage, crop residue, and manure
23 recycling) would increase N₂O emissions significantly, offsetting 75–310% of the C sequestered in terms of
24 CO₂ equivalence, while other practices such as cover crops can reduce N₂O emissions (Kaye and Quemada
25 2017).

26
27 The management of soil erosion could avoid a net emissions of 1.36– 3.67 GtCO₂ yr⁻¹ and create a sink of
28 0.44–3.67 GtCO₂ yr⁻¹ (*low confidence*) (Jacinthe and Lal 2001; Lal et al. 2004; Stallard 1998; Smith et al.
29 2001; Van Oost et al. 2007). The overall impact of erosion control on mitigation is context-specific and
30 uncertain at the global level, uncertain and the final fate of eroded material is still debated (Hoffmann et al.,
31 2013).

32
33 Biochar is produced by thermal decomposition of biomass in the absence of oxygen (pyrolysis) into a stable,
34 long-lived product like charcoal that is relatively resistant to decomposition (Lehmann et al. 2015) and which
35 can stabilise organic matter added to soil (Han Weng et al. 2017). Although charcoal has been used
36 traditionally by many cultures as a soil amendment, “modern biochar”, produced in facilities that control
37 emissions, is not widely used. The range of global potential of biochar is 0.03-6.6 GtCO₂-eq yr⁻¹ by 2050
38 including energy substitution, with 0.03-4.9 GtCO₂ yr⁻¹ for CDR only (*medium confidence*) (Griscom et al.
39 2017a; Hawken 2017; Paustian et al. 2016; Fuss et al. 2018; Lenton 2014, 2010; Powell and Lenton 2012a;
40 Woolf et al. 2010; Pratt and Moran 2010; Smith 2016; Roberts et al. 2010). An analysis in which biomass
41 supply constraints were applied to protect against food insecurity, loss of habitat and land degradation,
42 estimated *technical potential* abatement of 3.7–6.6 GtCO₂-eq yr⁻¹ (including 2.6–4.6 GtCO₂ yr⁻¹ carbon
43 stabilisation) (Woolf et al. 2010). Fuss et al. (2018) propose a range of 0.5–2 GtCO₂-eq yr⁻¹ as
44 the *sustainable potential* for negative emissions through biochar. (Griscom et al. 2017b) suggest a potential
45 of 1.0 GtCO₂ yr⁻¹ based on available residues. Biochar can provide additional climate change mitigation
46 benefits by decreasing nitrous oxide (N₂O) emissions from soil and reducing nitrogen fertiliser requirements
47 in agricultural soils (Borchard et al. 2019). Application of biochar to cultivated soils can darken the surface
48 and reduce its mitigation potential via decreases in surface albedo, but the magnitude of this effect depends
49 on soil moisture content, biochar application method and type of land use (*low confidence*) (Verheijen et al.
50 2013; Bozzi et al. 2015). Biochar is discussed in more detail in Chapter 4, Section 4.9.5.

51 52 **2.6.1.4 Land management in other ecosystems**

53 Protection and restoration of wetlands, peatlands and coastal habitats reduces net carbon loss (primarily from
54 sediment/soils) and provides continued or enhanced natural CO₂ removal (Chapter 4, section 4.9.4).
55 Reducing annual emissions from peatland conversion, draining and burning could mitigate 0.45-1.22 GtCO₂-
56 eq yr⁻¹ up to 2050 (*medium confidence*) (Hooijer et al. 2010; Griscom et al. 2017; Hawken 2017) and

1 peatland restoration 0.15 to 0.81 (*low confidence*) (Couwenberg et al. 2010; Griscom et al. 2017b). The
2 upper end from Griscom et al. (2017) represents a maximum sustainable potential (accounting for
3 biodiversity and food security safeguards) for rewetting and biomass enhancement. Wetland drainage and
4 rewetting was included as a flux category under the second commitment Period of the Kyoto protocol, with
5 significant management knowledge gained over the last decade (IPCC 2013c). However, there are high
6 uncertainties as to the carbon storage and flux rates, in particular the balance between CH₄ sources and CO₂
7 sinks (Spencer et al. 2016). Peatlands are sensitive to climate change which may increase carbon uptake by
8 vegetation and carbon emissions due to respiration, with the balance being regionally dependent (*high*
9 *confidence*). There is *low confidence* about the future peatland sink globally. Some peatlands have been
10 found to be resilient to climate change (Minayeva and Sirin 2012), but the combination of land use change
11 and climate change may make them vulnerable to fire (Sirin et al. 2011). While models show mixed results
12 for the future sink (Spahni et al. 2013; Chaudhary et al. 2017; Ise et al. 2008), a study that used extensive
13 historical data sets to project change under future warming scenarios found that the currently global peatland
14 sink could increase slightly until 2100 and decline thereafter (Gallego-Sala et al. 2018).

15
16 Reducing the conversion of coastal wetlands (mangroves, seagrass and marshes) could reduce emissions by
17 0.11–2.25 GtCO₂-eq yr⁻¹ by 2050 (*medium confidence*) (Pendleton et al. 2012; Griscom et al. 2017a; Howard
18 et al. 2017; Hawken 2017). Mangrove restoration can mitigate the release of 0.07 GtCO₂ yr⁻¹ through
19 rewetting (Crooks et al. 2011) and take up 0.02–0.84 GtCO₂ yr⁻¹ from biomass and soil enhancement
20 (Griscom et al. 2017b) (*medium confidence*). The ongoing benefits provided by mangroves as a natural
21 carbon sink can be nationally-important for Small Island Developing States (SIDS) and other countries with
22 extensive coastlines, based on estimates of high carbon sequestration rates per unit area (McLeod et al. 2011;
23 Duarte et al. 2013; Duarte 2017; Taillardat et al. 2018). There is only *medium confidence* in the effectiveness
24 of enhanced carbon uptake using mangroves, due to the many uncertainties regarding the response of
25 mangroves to future climate change (Jennerjahn et al. 2017); dynamic changes in distributions (Kelleway et
26 al. 2017) and other local-scale factors affecting long-term sequestration and climatic benefits (e.g., methane
27 release; Dutta et al. 2017). The climate mitigation potential of coastal vegetated habitats (mangrove forests,
28 tidal marshes and seagrasses) is considered in Chapter 5 of the IPCC Special Report on the Ocean,
29 Cryosphere and Climate Change (SROCC), in a wider ‘blue carbon’ context.

30 31 32 **2.6.1.5 Bioenergy and bioenergy with carbon capture and storage (BECCS)**

33 An introduction and overview of bioenergy and BECCS can be found in the Cross-Chapter Boxes: Cross-
34 Chapter Box 12: Traditional biomass use, Chapter 7; Cross-Chapter Box 7: Bioenergy and BECCS in
35 mitigation scenarios, Chapter 6. CCS technologies are discussed in SR15. The discussion below refers to
36 modern bioenergy only, (e.g., liquid biofuels for transport and the use of solid biofuels in combined heat and
37 power plants).

38
39 The mitigation potential of bioenergy coupled with CCS, i.e., BECCS, is estimated to be between 0.4 and
40 11.3 GtCO₂ yr⁻¹ (*medium confidence*) based on studies that directly estimate mitigation for BECCS (not
41 bioenergy) in units of CO₂ (not EJ) (McLaren 2012; Lenton 2014; Fuss et al. 2018; Turner et al. 2018b;
42 Lenton 2010; Koornneef et al. 2012; Powell and Lenton 2012b). SR15 reported a potential of 1-85 GtCO₂ yr⁻¹
43 ¹ which they noted could be narrowed to a range of 0.5 to 5 GtCO₂ yr⁻¹ when taking account of sustainability
44 aims (Fuss et al. 2018). The upper end of the SR15 range is considered as a theoretical potential. Previously,
45 the IPCC Special Report on Renewable Energy Sources concluded the technical potential of biomass supply
46 for energy (without BECCS) could reach 100-300 EJ yr⁻¹ by 2050, which would be 2-15 GtCO₂ yr⁻¹ (using
47 conversion factors 1 EJ = 0.02-0.05 GtCO₂ yr⁻¹ emission reduction, SR15). A range of recent studies
48 including sustainability or economic constraints estimate that 50-244 EJ (1 to 12 GtCO₂ yr⁻¹ using
49 conversion factors above) of bioenergy could be produced on 0.1-13 Mkm² of land (Fuss et al. 2018; Chan
50 and Wu 2015; Schueler et al. 2016; Wu et al. 2013; Searle and Malins 2015; Wu et al. 2019; Heck et al.
51 2018; Fritz et al. 2013) SR15 SPM).

52
53 There is *high confidence* that the most important factors determining future biomass supply for energy are
54 land availability and land productivity (Berndes et al. 2013; Creutzig et al. 2015a; Woods et al. 2015;
55 Daioglou et al. 2019). Estimates of marginal/degraded lands currently considered available for bioenergy
56 range from 3.2 to 14.0 Mkm², depending on the adopted sustainability criteria, land class definitions, soil

1 conditions, land mapping method and environmental and economic considerations (Campbell et al. 2008;
2 Cai et al. 2011; Lewis and Kelly 2014).

3
4 Bioenergy production systems can lead to net emissions in the short term that can be “paid-back” over time,
5 with multiple harvest cycles and fossil fuel substitution, unlike fossil carbon emissions (Campbell et al.
6 2008; Cai et al. 2011; Lewis and Kelly 2014; De Oliveira Bordonal et al. 2015). Stabilising bioenergy crops
7 in previous high carbon forestland or peatland results in high emissions of carbon that may take from
8 decades to more than a century to be re-paid in terms of net CO₂ emission savings from replacing fossil
9 fuels, depending on previous forest carbon stock, bioenergy yields, and displacement efficiency (Elshout et
10 al. 2015; Harper et al. 2018; Daioglou et al. 2017). In the case of bioenergy from managed forests, the
11 magnitude and timing of the net mitigation benefits is controversial, as it varies with differences due to local
12 climate conditions, forest management practice, fossil fuel displacement efficiency and methodological
13 approaches (Hudiburg et al. 2011; Berndes et al. 2013; Guest et al. 2013; Lamers and Junginger 2013;
14 Cherubini et al. 2016; Cintas et al. 2017; Laurance et al. 2018; Valade et al. 2018; Baker et al. 2019).
15 Suitable bioenergy crops can be integrated in agricultural landscapes to reverse ecosystem carbon depletion
16 (Creutzig et al. 2015a; Robertson et al. 2017; Vaughan et al. 2018; Daioglou et al. 2017). Cultivation of short
17 rotation woody crops and perennial grasses on degraded land or cropland previously used for annual crops
18 typically accumulate carbon in soils due to their deep root systems (Don et al. 2012; Robertson et al. 2017).
19 The use of residues and organic waste as bioenergy feedstock can mitigate land use change pressures
20 associated with bioenergy deployment, but residues are limited and the removal of residues that would
21 otherwise be left on the soil could lead soil degradation (Chum et al. 2011; Liska et al. 2014; Monforti et al.
22 2015; Zhao et al. 2015; Daioglou et al. 2016).

23
24 The steps required to cultivate, harvest, transport, process and use biomass for energy generate emissions of
25 GHGs and other climate pollutants (Chum et al. 2011; Creutzig et al. 2015b; Staples et al. 2017; Daioglou et
26 al. 2019). Life-cycle GHG emissions of modern bioenergy alternatives are usually lower than those for fossil
27 fuels (*robust evidence, medium agreement*) (Chum et al. 2011; Creutzig et al. 2015b). The magnitude of
28 these emissions largely depends on location, (e.g. soil quality, climate), prior land use, feedstock used (e.g.,
29 residues, dedicated crops, algae), land use practice (e.g., soil management, fertiliser use), biomass transport
30 (distances and transport modes), the bioenergy conversion pathway and product (e.g., wood pellets, ethanol).
31 Use of conventional food and feed crops as a feedstock generally provides the highest bioenergy yields per
32 hectare, but also causes more GHG emissions per unit energy compared to agriculture residues, biomass
33 from managed forests, and lignocellulosic crops such as short-rotation coppice and perennial grasses (Chum
34 et al. 2011; Gerbrandt et al. 2016). This is due to the application of fertilisers and other inputs (Oates et al.
35 2016; Rowe et al. 2016; Lai et al. 2017; Robertson et al. 2017).

36
37 Bioenergy from dedicated crops are in some cases held responsible for GHG emissions resulting from
38 indirect land use change (iLUC), that is the bioenergy activity may lead to displacement of agricultural or
39 forest activities into other locations, driven by market-mediated effects. Other mitigation options may also
40 cause iLUC. At a global level of analysis, indirect effects are not relevant because all land-use emissions are
41 direct. iLUC emissions are potentially more significant for crop-based feedstocks such as corn, wheat and
42 soybean, than for advanced biofuels from lignocellulosic materials (Chum et al. 2011; Wicke et al. 2012;
43 Valin et al. 2015; Ahlgren and Di Lucia 2014). Estimates of emissions from iLUC are inherently uncertain,
44 are widely debated in the scientific community, and are highly dependent on modelling assumptions, such as
45 supply/demand elasticities, productivity estimates, incorporation or exclusion of emission credits for
46 coproducts, scale of biofuel deployment (Rajagopal and Plevin 2013; Finkbeiner 2014; Kim et al. 2014;
47 Zilberman 2017). In some cases, iLUC effects are estimated to result in emission reductions. For example,
48 market-mediated effects of bioenergy in North America showed potential for increased carbon stocks by
49 inducing conversion of pasture or marginal land to forestland (Cintas et al. 2017; Duden et al. 2017; Dale et
50 al. 2017; Baker et al. 2019). There is a wide range of variability in iLUC values for different types of
51 biofuels, from -75 to +55 g CO₂ MJ⁻¹ (Ahlgren and Di Lucia 2014; Valin et al. 2015; Plevin et al. 2015;
52 Taheripour and Tyner 2013; Bento and Klotz 2014). There is low confidence in attribution of emissions from
53 iLUC to bioenergy.

54
55 Bioenergy deployment can have large biophysical effects on regional climate, with the direction and
56 magnitude of the impact depending on the type of bioenergy crop, previous land use and seasonality (*limited*

1 *evidence, medium agreement*). A study of two alternative future bioenergy scenarios using 15 Mkm² of
2 intensively used managed land or conversion of natural areas showed a nearly neutral effect on surface
3 temperature at global levels (considering biophysical effects and CO₂ and N₂O fluxes from land but not
4 substitution effects), although there were significant seasonal and regional differences (Kicklighter et al.
5 2013). Modelling studies on biofuels in the US found the switch from annual crops to perennial bioenergy
6 plantations like Miscanthus in the US could lead to regional cooling due to increases in evapotranspiration
7 and albedo (Georgescu et al. 2011; Harding et al. 2016), with perennial bioenergy crop expansion over
8 suitable abandoned and degraded farmlands causing near-surface cooling up to 5°C during the growing
9 season (Wang et al. 2017b). Similarly, growing sugarcane on existing cropland in Brazil cools down the
10 local surface during daytime conditions up to -1°C, but warmer conditions occurs if sugar cane is deployed at
11 the expense of natural vegetation (Brazilian Cerrado) (Loarie et al. 2011). In general, bioenergy crops (as for
12 all crops) induce a cooling of ambient air during the growing season, but after harvest the decrease in
13 evapotranspiration can induce warming (Harding et al. 2016; Georgescu et al. 2013; Wang et al. 2017b).
14 Bioenergy crops were found to cause increased isoprene emissions in a scenario where 0.69 Mkm² of oil
15 palm for biodiesel in the tropics and 0.92 Mkm² of short rotation coppice (SRC) in the mid-latitudes were
16 planted, but effects on global climate were negligible (Ashworth et al. 2012).

17 18 **2.6.1.6 Enhanced weathering**

19 Weathering is the natural process of rock decomposition via chemical and physical processes in which CO₂
20 is removed from the atmosphere and converted to bicarbonates and/or carbonates (IPCC 2005). Formation of
21 calcium carbonates in the soil provides a permanent sink for mineralised organic carbon (Manning 2008;
22 Beerling et al. 2018). Mineral weathering can be enhanced through grinding up rock material to increase the
23 surface area, and distributing it over land to provide carbon removals of 0.5–4.0 GtCO₂ yr⁻¹ (*medium*
24 *confidence*) (Beerling et al. 2018; Lenton 2010; Smith et al. 2016a; Taylor et al. 2016). While the
25 geochemical potential is quite large, agreement on the technical potential is low due to a variety of unknown
26 parameters and of limits such as rates of mineral extraction, grinding, delivery, and challenges with scaling
27 and deployment.

28 29 **2.6.1.7 Demand management in the food sector (diet change, waste reduction)**

30 Demand-side management has the potential for climate change mitigation via reducing emissions from
31 production, switching to consumption of less emission intensive commodities, and making land available for
32 carbon dioxide removal (see Chapter 5 Section 5.5.2). Reducing food losses and waste increases the overall
33 efficiency of food value chains (less land and inputs needed) along the entire supply chain, and has the
34 potential to mitigate 0.8-4.5 GtCO₂-eq yr⁻¹ (*high confidence*) (Chapter 5 section 5.5.2.5, Bajželj et al. 2014;
35 Dickie et al 2014; Hawken 2017; Hiç et al. 2016).

36
37 Shifting to diets that are lower in emissions-intensive foods like beef delivers a mitigation potential of 0.7-
38 8.0 GtCO₂-eq yr⁻¹ (*high confidence*) (Bajželj et al. 2014; Dickie et al. 2014; Herrero et al. 2016; Hawken
39 2017; Springmann et al. 2016; Tilman and Clark 2014; Hedenus et al. 2014; Stehfest et al. 2009) with most
40 of the higher end estimates (> 6 GtCO₂-eq yr⁻¹) based on veganism, vegetarianism or very low ruminant meat
41 consumption) (see Chapter 5, Section 5.5.2). In addition to direct mitigation gains, decreasing meat
42 consumption, primarily of ruminants, and reducing wastes further reduces water use, soil degradation,
43 pressure on forests, land used for feed potentially freeing up land for mitigation (Tilman and Clark 2014)
44 (see chapters 5 and 6). Additionally, consumption of locally produced food, shortening the supply chain, can
45 in some cases minimise food loss, contribute to food security, and reduce GHG emissions associated with
46 energy consumption and food loss (see Chapter 5 Section 5.5.2.6).

47 48 **2.6.2 Integrated pathways for climate change mitigation**

49
50 Land-based response options have the potential to interact, resulting in additive effects (e.g., climate co-
51 benefits) or negating each other (e.g., through competition for land), they also interact with mitigation
52 options in other sectors (such as energy or transport), thus they need to be assessed collectively under
53 different climate mitigation targets and in combination with other sustainability goals (Popp et al. 2017;
54 Obersteiner et al. 2016; Humpenöder et al. 2018). Integrated Assessment Models (IAMs) with distinctive
55 land-use modules are the basis for the assessment of mitigation pathways as they combine insights from
56 various disciplines in a single framework and cover the largest sources of anthropogenic GHG emissions

1 from different sectors (see also SR15 Chapter 2 and technical Annex for more details). IAMs consider a
2 limited, but expanding, portfolio of land-based mitigation options. Furthermore, the inclusion and detail of a
3 specific mitigation measure differs across IAMs and studies (see also SR15 and Chapter 6). For example, the
4 IAM scenarios based on the Shared Socio-economic Pathways (SSPs) (Riahi et al. 2017)(see more details on
5 the SSPs in Cross-Chapter Box 1: Scenarios, Chapter 1) include possible trends in agriculture and land use
6 for five different socioeconomic futures, but cover a limited set of land-based mitigation options: dietary
7 changes, higher efficiency in food processing (especially in livestock production systems), reduction of food
8 waste, increasing agricultural productivity, methane reductions in rice paddies, livestock and grazing
9 management for reduced methane emissions from enteric fermentation, manure management, improvement
10 of N-efficiency, 1st generation biofuels, reduced deforestation, afforestation, 2nd generation bioenergy crops
11 and BECCS (Popp et al. 2017). However, many “natural climate solutions” (Griscom et al. 2017b), such as
12 forest management, rangeland management, soil carbon management or wetland management, are not
13 included in most of these scenarios. In addition, most IAMs neglect the biophysical effects of land-use such
14 as changes in albedo or evapotranspiration with few exceptions (Kreidenweis et al. 2016).

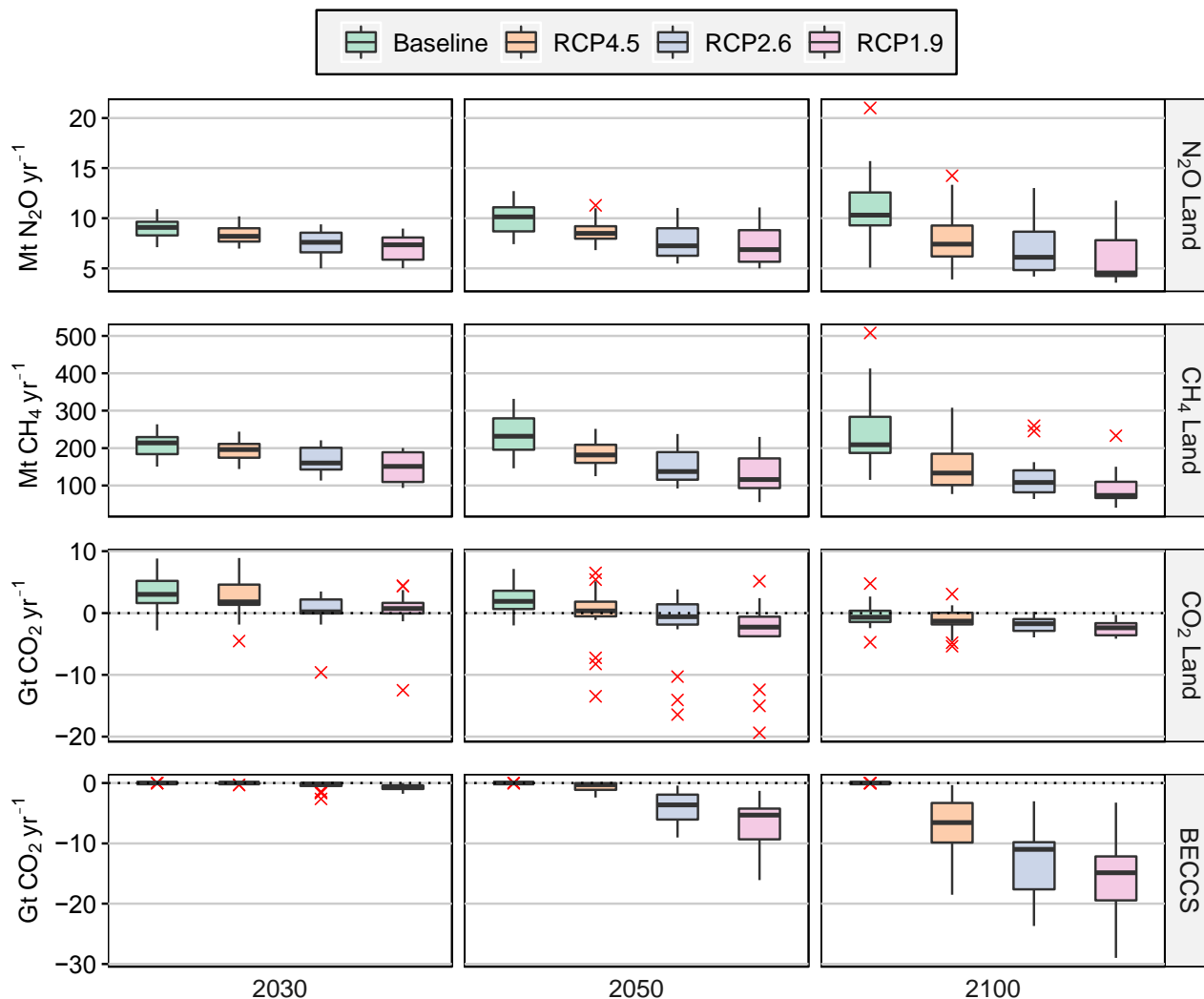
15
16 Mitigation pathways, based on IAMs, are typically designed to find the least cost pathway to achieve a pre-
17 defined climate target (Riahi et al. 2017). Such cost-optimal mitigation pathways, especially in RCP2.6
18 (broadly a 2°C target) and 1.9 scenarios (broadly a 1.5°C target), project GHG emissions to peak early in the
19 21st century, strict GHG emission reduction afterwards and, depending on the climate target, net carbon
20 dioxide removal (CDR) from the atmosphere in the second half of the century (see Chapter 2 of SR15,
21 (Tavoni et al. 2015; Riahi et al. 2017). In most of these pathways, land use is of great importance because of
22 its mitigation potential as discussed in section 2.7.1: these pathways are based on the assumptions that large-
23 scale afforestation and reforestation removes substantial amounts of CO₂ from the atmosphere; biomass
24 grown on cropland or from forestry residues can be used for energy generation or BECCS substituting fossil
25 fuel emissions and generating CDR; non-CO₂ emissions from agricultural production can be reduced, even
26 under improved agricultural management (Popp et al. 2017; Rogelj et al. 2018a; Van Vuuren et al. 2018,
27 Frank et al. 2018).

28
29 From the IAM scenarios available to this assessment, a set of feasible mitigation pathways has been
30 identified which is illustrative of the range of possible consequences on land use and GHG emissions
31 (presented in this chapter) and sustainable development (see Chapter 6). Thus, the IAM scenarios selected
32 here vary due to underlying socio-economic and policy assumptions, mitigation options considered, long-
33 term climate goal, the level of inclusion of other sustainability goals (such as land and water restrictions for
34 biodiversity conservation or food production), and models by which they are generated.

35
36 In the baseline case without climate change mitigation, global CO₂ emissions from land-use change decrease
37 over time in most scenarios due to agricultural intensification and decreases in demand for agricultural
38 commodities – some turning even negative by the end of the century due to abandonment of agricultural land
39 and associated carbon uptake through vegetation regrowth. Median global CO₂ emissions from land-use
40 change across 5 SSPs and 5 IAMs decrease throughout the 21st century: 3, 1.9 and -0.7 GtCO₂ yr⁻¹ in 2030,
41 2050 and 2100 respectively (Figure 2.25). In contrast, CH₄ and N₂O emissions from agricultural production
42 remain rather constant throughout the 21st century (CH₄: 214, 231.7 and 209.1 Mt CH₄ yr⁻¹ in 2030, 2050
43 and 2100 respectively; N₂O: 9.1, 10.1 and 10.3 Mt N₂O yr⁻¹ in 2030, 2050 and 2100 respectively).

44
45 In the mitigation cases (RCP4.5, RCP2.6 and RCP1.9), most of the scenarios indicate strong reductions in
46 CO₂ emissions due to i) reduced deforestation and ii) carbon uptake due to afforestation. However, CO₂
47 emissions from land use can occur in some mitigation scenarios as a result of weak land-use change
48 regulation (Fujimori et al. 2017; Calvin et al. 2017) or displacement effects into pasture land caused by high
49 bioenergy production combined with forest protection only (Popp et al. 2014). The level of carbon dioxide
50 removal globally (median value across SSPs and IAMs) increases with the stringency of the climate target
51 (RCP4.5, RCP2.6 and RCP1.9) for both afforestation (-1.3, -1.7 and -2.4 GtCO₂ yr⁻¹ in 2100) and BECCS (-
52 6.5, -11 and -14.9 GtCO₂ yr⁻¹ in 2100; see also Cross-Chapter Box 7: Bioenergy and BECCS in mitigation
53 scenarios, Chapter 6). In the mitigation cases (RCP4.5, RCP2.6 and RCP1.9), CH₄ and N₂O emissions are
54 remarkably lower compared to the baseline case (CH₄: 133.2, 108.4 and 73.5 Mt CH₄ yr⁻¹ in 2100; N₂O: 7.4,
55 6.1 and 4.5 Mt N₂O yr⁻¹ in 2100; see previous paragraph for CH₄ and N₂O emissions in the baseline case).
56 The reductions in the mitigation cases are mainly due to improved agricultural management such as

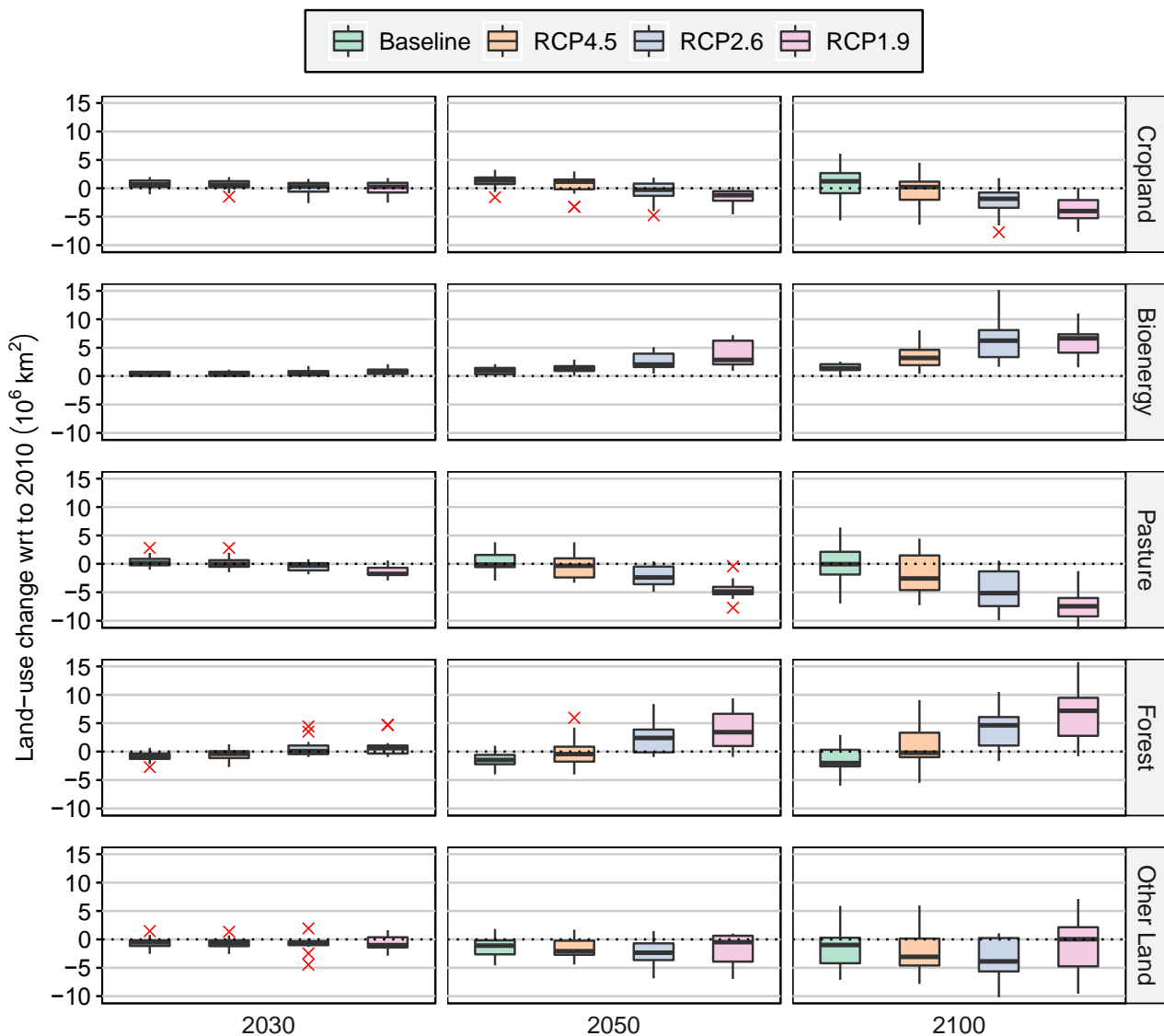
1 improved nitrogen fertiliser management, improved water management in rice production, improved manure
 2 management by for example covering of storages or adoption of biogas plants, better herd management and
 3 better quality of livestock through breeding and improved feeding practices. In addition, dietary shifts away
 4 from emission-intensive livestock products also lead to decreased CH₄ and N₂O emissions especially in
 5 RCP2.6 and RCP1.9 scenarios. However, high levels of bioenergy production can result in increased N₂O
 6 emissions due to nitrogen fertilisation of dedicated bioenergy crops.
 7



8
 9
 10 **Figure 2.25 Land-based global GHG emissions and removals in 2030, 2050 and 2100 for Baseline, RCP4.5,**
 11 **RCP2.6 and RCP1.9 based on the Shared Socioeconomic Pathways (SSP) (Popp et al. 2017; Rogelj et al.**
 12 **2018; Riahi et al. 2017). Data is from an update of the IAMC Scenario Explorer developed for the SR15**
 13 **(Huppmann et al. 2018; Rogelj et al. 2018). Boxplots (Tukey style) show median (horizontal line),**
 14 **interquartile range (IQR box) and the range of values within 1.5 x IQR at either end of the box (vertical**
 15 **lines) across 5 SSPs and across 5 IAMs. Outliers (red crosses) are values greater than 1.5 x IQR at either**
 16 **end of the box. The categories CO₂ Land, CH₄ Land and N₂O Land include GHG emissions from land-use**
 17 **change and agricultural land use (including emissions related to bioenergy production). In addition, the**
 18 **category CO₂ Land includes negative emissions due to afforestation. BECCS reflects the CO₂ emissions**
 19 **captured from bioenergy use and stored in geological deposits.**
 20
 21
 22

23 Such high levels of carbon dioxide removal through mitigation options that require land conversion (BECCS
 24 and afforestation) shape the land system dramatically (Figure 2.26). Across the different RCPs, SSPs and
 25 IAMs median change of global forest area throughout the 21st century ranges from about -0.2 to +7.2 Mkm²
 26 between 2010 and 2100, and agricultural land used for 2nd generation bioenergy crop production ranges from

1 about 3.2–6.6 Mkm² in 2100 (Popp et al. 2017; Rogelj et al. 2018). Land requirements for bioenergy and
 2 afforestation for a RCP1.9 scenario are higher than for a RCP2.6 and especially a RCP4.5 mitigation
 3 scenario. As a consequence of the expansion of mainly land-demanding mitigation options, global pasture
 4 land is reduced in most mitigation scenarios much stronger compared to baseline scenarios (median
 5 reduction of 0, 2.6, 5.1 and 7.5 Mkm² between 2010 and 2100 in Baseline, RCP4.5, RCP2.6 and RCP1.9
 6 respectively). In addition, cropland for food and feed production decreases with the stringency of the climate
 7 target (+1.2, +0.2, -1.8 and -4 Mkm² in 2100 compared to 2010 in Baseline, RCP4.5, RCP2.6 and RCP1.9
 8 respectively). These reductions in agricultural land for food and feed production are facilitated by
 9 agricultural intensification on agricultural land and in livestock production systems (Popp et al. 2017) but
 10 also by changes in consumption patterns (Fujimori et al. 2017; Frank et al. 2017b). The pace of projected
 11 land-use change over the coming decades in ambitious mitigation scenarios goes well beyond historical
 12 changes in some instances (Turner et al. 2018c), see also SR15). This raises issues for societal acceptance,
 13 and distinct policy and governance for avoiding negative consequences for other sustainability goals
 14 (Humpenöder et al. 2018; Obersteiner et al. 2016; Calvin et al. 2014), see Chapter 6 and 7).
 15
 16



17
 18 **Figure 2.26** Global change of major land cover types by 2030, 2050 and 2100 relative to 2010 for Baseline,
 19 RCP4.5, RCP2.6 and RCP1.9 based on the Shared Socioeconomic Pathways (SSP) (Popp et al. 2017; Rogelj et
 20 al. 2018; Riahi et al. 2017). Data is from an update of the IAMC Scenario Explorer developed for the SR15
 21 (Huppmann et al. 2018; Rogelj et al. 2018). Boxplots (Tukey style) show median (horizontal line),
 22 interquartile range IQR (box) and the range of values within 1.5 x IQR at either end of the box (vertical lines)
 23 across 5 SSPs and across 5 IAMs. Outliers (red crosses) are values greater than 1.5 x IQR at either end of the
 24 box. In 2010, total land cover at global scale was estimated 15-16 Mkm² for cropland, 0-0.14 Mkm² for

1 **bioenergy, 30-35 Mkm² for pasture and 37-42 Mkm² for forest, across the IAMs that reported SSP pathways**
2 **(Popp et al. 2017).**
3

4 Different mitigation strategies can achieve the net emissions reductions that would be required to follow a
5 Pathway that limits global warming to 2°C or 1.5°C, with very different consequences on the land system.
6

7 **Figure 2.27** shows six alternative pathways (archetypes) for achieving ambitious climate targets (RCP2.6
8 and RCP1.9) highlighting land-based strategies and GHG emission. All pathways are assessed by different
9 models but are all based on the Shared Socioeconomic Pathway 2 (SSP2) (Riahi et al. 2017), with all based
10 on an RCP 1.9 mitigation pathway expect for Pathway 1, which is RCP2.6. All scenarios show land-based
11 negative emissions but the amount varies across pathways, as do the relative contributions of different land-
12 based Carbon Dioxide Removal (CDR) options, such as afforestation/reforestation and bioenergy with
13 carbon capture and storage (BECCS).
14

15 Pathway 1 RCP2.6 “Portfolio” (Fricko et al. 2017) shows a strong near-term decrease of CO₂ emissions from
16 land-use change, mainly due to reduced deforestation, as well as slightly decreasing N₂O and CH₄ emissions
17 after 2050 from agricultural production due to improved agricultural management and dietary shifts away
18 from emissions-intensive livestock products. However, in contrast to CO₂ emissions, which turn net-negative
19 around 2050 due to afforestation/reforestation, CH₄ and N₂O emissions persist throughout the century due to
20 difficulties of eliminating these residual emissions based on existing agricultural management methods
21 (Stevanović et al. 2017; Frank et al. 2017b). In addition to abating land-related GHG emissions as well as
22 increasing the terrestrial sink, this example also shows the importance of the land sector in providing
23 biomass for BECCS and hence CDR in the energy sector. In this scenario, annual BECCS-based CDR is
24 about 3-times higher than afforestation-based CDR in 2100 (-11.4 and -3.8 GtCO₂ yr⁻¹ respectively).
25 Cumulative CDR throughout the century amounts to -395 GtCO₂ for BECCS and -73 GtCO₂ for
26 afforestation. Based on these GHG dynamics, the land sector turns GHG emission neutral in 2100. However,
27 accounting also for BECCS-based CDR taking place in the energy sector but with biomass provided by the
28 land sector turns the land sector GHG emission neutral already in 2060, and significantly net-negative by the
29 end of the century.
30

31 Pathway 2 RCP1.9 “Increased Ambition” (Rogelj et al. 2018) has dynamics of land-based GHG emissions
32 and removals that are very similar to those in Pathway 1 (RCP2.6) but all GHG emission reductions as well
33 as afforestation/reforestation and BECCS-based CDR start earlier in time at a higher rate of deployment.
34 Cumulative CDR throughout the century amounts to -466 GtCO₂ for BECCS and -117 GtCO₂ for
35 afforestation.
36

37 Pathway 3 RCP 1.9 “Only BECCS”, in contrast to Pathway 2, includes only BECCS-based CDR (Kriegler
38 et al. 2017). In consequence, CO₂ emissions are persistent much longer, predominantly from indirect land-
39 use change due to large-scale bioenergy cropland expansion into non-protected natural areas (Popp et al.
40 2017; Calvin et al. 2014). While annual BECCS CDR rates in 2100 are similar to Pathway 1 and 2 (-15.9
41 GtCO₂ yr⁻¹), cumulative BECCS-based CDR throughout the century is much larger (-944 GtCO₂).
42

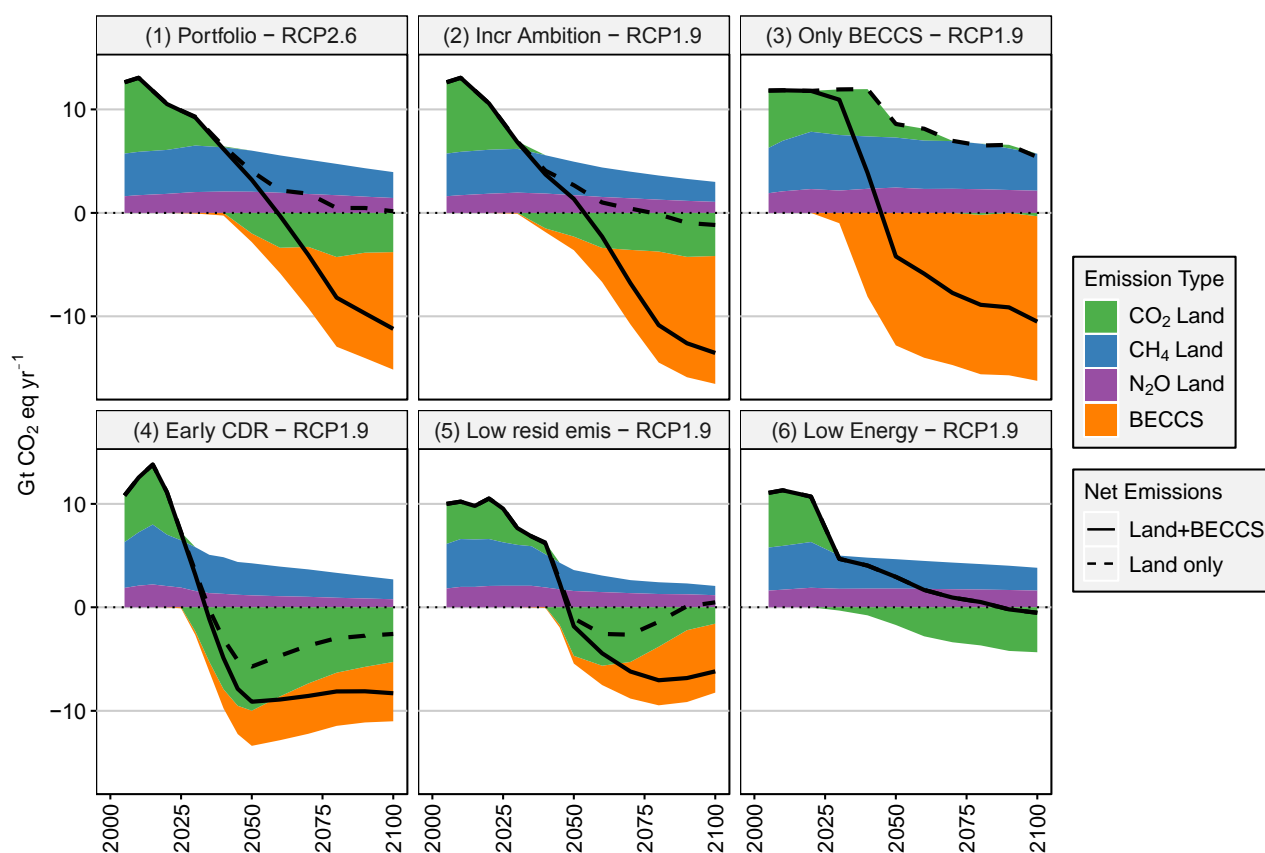
43 Pathway 4 RCP1.9 “Early CDR” (Bertram et al. 2018) indicates that a significant reduction in the later
44 century in the BECCS-related CDR as well as CDR in general can be achieved with earlier and mainly
45 terrestrial CDR, starting already in 2030. In this scenario, terrestrial CDR is based on afforestation but could
46 also be supported by soil organic carbon sequestration (Paustian et al. 2016) or other natural climate
47 solutions such as rangeland or forest management (Griscom et al. 2017b). This scenario highlights the
48 importance of the timing for CDR-based mitigation pathways (Obersteiner et al. 2016). As a result of near-
49 term and mainly terrestrial CDR deployment, cumulative BECCS-based CDR throughout the century is
50 limited to -300 GtCO₂, while cumulative afforestation-based CDR amounts to -428 GtCO₂.
51

52 In Pathway 5 RCP1.9 “Low residual emissions” (van Vuuren et al. 2018), land-based mitigation is driven by
53 stringent enforcement of measures and technologies to reduce end-of-pipe non-CO₂ emissions and by
54 introduction of in-vitro (cultured) meat, reducing residual N₂O and CH₄ emissions from agricultural
55 production. In consequence, much lower amounts of CDR from afforestation and BECCS are needed with
56 much later entry points to compensate for residual emissions. Cumulative CDR throughout the century

1 amounts to -252 GtCO₂ for BECCS and -128 GtCO₂ for afforestation. Therefore, total cumulative land-based
 2 CDR in Pathway 5 is substantially lower compared to Pathways 2-4 (380 GtCO₂).
 3

4 Finally, Pathway 6 RCP1.9 “Low Energy” (Grubler et al. 2018) – equivalent to pathway LED in SR15 –
 5 indicates the importance of other sectoral GHG emission reductions for the land sector. In this example,
 6 rapid and early reductions in energy demand and associated drops in energy-related CO₂ emissions, limit
 7 overshoot and decrease the requirements for negative emissions technologies, especially for land-demanding
 8 CDR such as biomass production for BECCS and afforestation. While BECCS is not used at all in Pathway 6,
 9 cumulative CDR throughout the century for afforestation amounts to -124 GtCO₂.
 10

11 Besides their consequences on mitigation pathways and land consequences, those archetypes can also affect
 12 multiple other sustainable development goals that provide both challenges and opportunities for climate
 13 action (see Chapter 6).
 14
 15



16
 17
 18 **Figure 2.27 Evolution and break down of global land-based GHG emissions and removals under six**
 19 **alternative mitigation pathways, which illustrate the differences in timing and magnitude of land-based**
 20 **mitigation approaches including afforestation and BECCS. All pathways are based on different IAM**
 21 **realisations of SSP2. Pathway 1 is based on RCP 2.6, while all other pathways are based on RCP 1.9.**
 22 **Pathway 1: MESSAGE-GLOBIOM (Fricko et al. 2017); Pathway 2: MESSAGE-GLOBIOM (Rogelj et**
 23 **al. 2018); Pathway 3: REMIND-MAGPIE (Kriegler et al. 2017); Pathway 4: REMIND-MAGPIE**
 24 **(Bertram et al. 2018); Pathway 5: IMAGE (van Vuuren et al. 2018); Pathway 6: MESSAGE-GLOBIOM**
 25 **(Grubler et al. 2018). Data is from an update of the IAMC Scenario Explorer developed for the SR15**
 26 **(Rogelj et al. 2018). The categories CO₂ Land, CH₄ Land and N₂O Land include GHG emissions from**
 27 **land-use change and agricultural land use (including emissions related to bioenergy production). In**
 28 **addition, the category CO₂ Land includes negative emissions due to afforestation. BECCS reflects the**
 29 **CO₂ emissions captured from bioenergy use and stored in geological deposits. Solid lines show the net**
 30 **effect of all land-based GHG emissions and removals (CO₂ Land, CH₄ Land, N₂O Land and BECCS),**
 31 **while dashed lines show the net effect excluding BECCS. CH₄ and N₂O emissions are converted to CO₂-**
 32 **eq using GWP factors of 28 and 265 respectively.**
 33

2.6.3 The contribution of response options to the Paris Agreement

The previous sections indicated how land based response options have the potential to contribute to the Paris Agreement, not only through reducing anthropogenic emissions but also for providing anthropogenic sinks that can contribute to “...a balance between anthropogenic emissions by sources and removals by sinks of greenhouse gases in the second half of this century ...” (Paris Agreement, Article 4). The balance applies globally, and relates only to greenhouse gases, not aerosols (Section 2.4) or biophysical effects (Section 2.5).

The Paris Agreement includes an Enhanced Transparency Framework, to track countries’ progress towards achieving their individual targets (i.e., NDCs), and a Global Stocktake (every five years starting in 2023), to assess the countries’ collective progress towards the long-term goals of the Paris Agreement. The importance of robust and transparent definitions and methods (including the approach to separating anthropogenic from natural fluxes) (Fuglestad et al. 2018) and the needs for reconciling country GHG inventories and models (Grassi et al. 2018a) was highlighted in 2.3 in relation to estimating emissions. Issues around estimating mitigation is also key to transparency and credibility and is part of the Paris Rulebook.

The land sector is expected to deliver up to 25% of GHG mitigation pledged by countries by 2025-2030 in their NDCs, based on early assessments of “Intended” NDCs submitted ahead of the Paris Agreement and updates immediately after (*low confidence*) (Grassi et al. 2017; Forsell et al. 2016). While most NDCs submitted to date include commitments related to the land sector, they vary with how much information is given and the type of target, with more ambitious targets for developing countries often being “conditional” on support and climate finance. Some do not specify the role of AFOLU but include it implicitly as part of economy-wide pledges (e.g. reducing total emission or emission intensity), a few mention multi-sectoral mitigation targets which include AFOLU in a fairly unspecified manner. Many NDCs include specific AFOLU response options, with most focused on the role of forests, a few included soil carbon sequestration or agricultural mitigation, few explicitly mention bioenergy (e.g., Cambodia, Indonesia and Malaysia), but this could be implicitly included with reduced emission in energy sector through fuel substitution (see Cross-Chapter Box 7: Bioenergy and BECCS in mitigation scenarios, Chapter 6, for discussion on cross sector flux reporting). The countries indicating AFOLU mitigation most prominently were Brazil and Indonesia, followed by other countries focusing either on avoiding carbon emissions (e.g., Ethiopia, Gabon, Mexico, DRC, Guyana and Madagascar) or on promoting the sink through large afforestation programs (e.g., China, India) (Grassi et al. 2017).

Figure 2.28 shows the CO₂ mitigation potential of NDCs compared to historical fluxes from LULUCF⁴. It shows future fluxes based on current policies in place and on country-stated Business As Usual (BAU) activities (these are different from current policies as many countries are already implementing policies that they do not include as part of their historical business-as-usual baseline) (Grassi et al. 2017). Under implementation of unconditional pledges, the net LULUCF flux in 2030 has been estimated to be a sink of $-0.41 \pm 0.68 \text{ GtCO}_2 \text{ yr}^{-1}$, which increases to $-1.14 \pm 0.48 \text{ GtCO}_2 \text{ yr}^{-1}$ in 2030 with conditional activities. This compares to net LULUCF in 2010 calculated from the GHG Inventories of $0.01 \pm 0.86 \text{ GtCO}_2 \text{ yr}^{-1}$ (Grassi et al. 2017). Forsell et al. (2016) similarly find a reduction in 2030 compared to 2010 of $0.5 \text{ GtCO}_2 \text{ yr}^{-1}$ (range: 0.2–0.8) by 2020 and $0.9 \text{ GtCO}_2 \text{ yr}^{-1}$ (range: 0.5–1.3) by 2030 for unconditional and conditional cases.

The approach of countries to calculating the LULUCF contribution towards the NDC varies, with implications for comparability and transparency. For example, by following the different approaches used to include LULUCF in country NDCs, (Grassi et al. 2017) found a 3-fold difference in estimated mitigation: $1.2\text{--}1.9 \text{ GtCO}_2\text{-eq yr}^{-1}$ when 2030 expected emissions are compared to 2005 emissions; $0.7\text{--}1.4 \text{ GtCO}_2\text{-eq yr}^{-1}$ when 2030 emissions are compared to reference scenarios based on current policies or $2.3\text{--}3.0 \text{ GtCO}_2\text{-eq yr}^{-1}$ when compared to BAU, and $3.0\text{--}3.8 \text{ GtCO}_2\text{-eq yr}^{-1}$ when based on using each countries’ approach to calculation stated in the NDC (i.e., when based on a mix of country approaches, using either past years or BAU projections as reference)

⁴ FOOTNOTE: CO₂ fluxes due to Land Use, Land-Use Change and Forestry, i.e. not including the part of AFOLU fluxes that are from agriculture

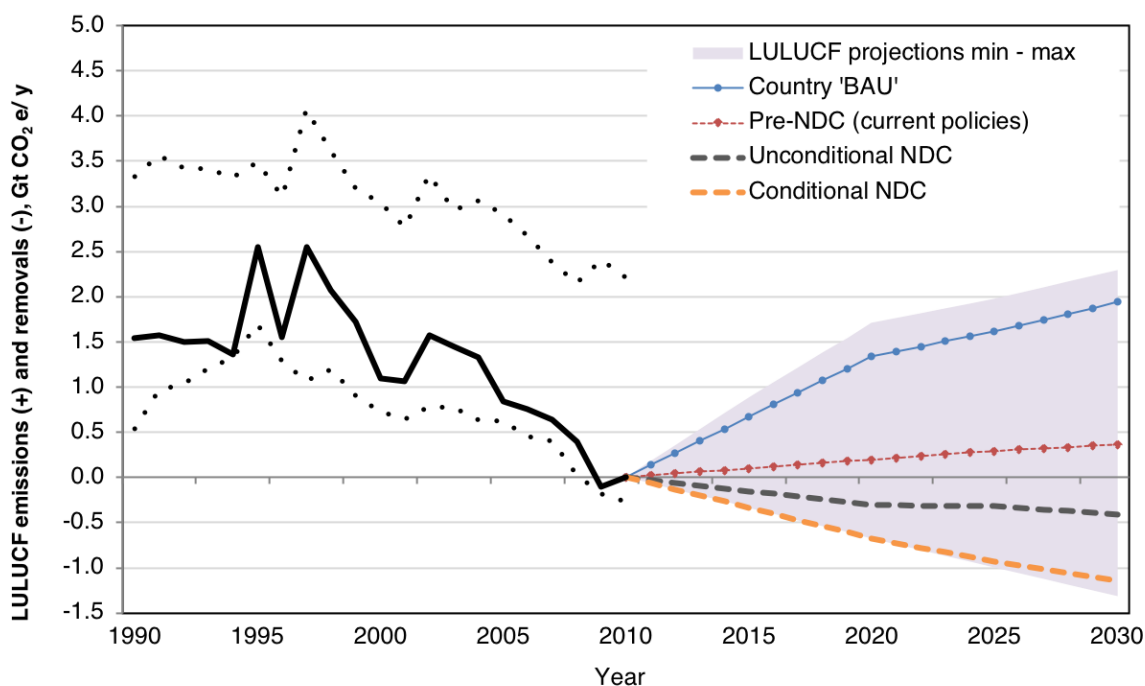


Figure 2.28 Global Land Use, Land Use Change and Forestry (LULUCF) net greenhouse gas flux for the historical period and future scenarios based on analyses of countries' Nationally Determined Contributions (NDCs). The LULUCF historical data (black solid line) reflect the following countries' documents (in order of priority): data submitted to UNFCCC (NDCs⁵, 2015 GHG Inventories⁶, recent National Communications^{7, 8}); other official countries' documents; FAO-based datasets, i.e. FAO-FRA for forest (Tian et al. 2015) as elaborated by (Federici et al. 2015) and FAOSTAT for non-forest land use emissions (FAO 2015a). The future four scenarios reflect official countries' information, mostly INDCs or updated NDCs available at the time of the analysis (Feb 2016), complemented by Biennial Update Reports⁹ and National Communications, and show: the Business as Usual (BAU) scenario as defined by the country; the trend based on pre-NDC levels of activity (current policies in place in countries); and the unconditional NDC and conditional NDC scenarios. The shaded area indicates the full range of countries' available projections (min-max), expressing the available countries' information on uncertainties beyond the specific scenarios shown. The range of historical country datasets (dotted lines) reflects differences between alternative selections of country sources, i.e. GHG inventories for developed countries complemented by FAO-based datasets (upper range) or by data in National Communications (lower range) for developing countries

In Exploring the effectiveness of the NDCs, SR15 concluded “Estimates of global average temperature increase are 2.9–3.4°C above preindustrial levels with a greater than 66% probability by 2100 (Roberts et al. 2006; Rogelj et al. 2016), under a full implementation of unconditional NDCs and a continuation of climate action similar to that of the NDCs. In order to achieve 1.5°C or 2°C pathways, this shortfall would imply the need for submission (and achievement) of more ambitious NDCs, and plan for a more rapid transformation of their national energy, industry, transport, and land use sectors (Peters and Geden 2017; Millar et al. 2017; Rogelj et al. 2016).

⁵ FOOTNOTE: UNFCCC. INDCs as communicated by Parties, <http://www4.unfccc.int/submissions/indc/Submission%20Pages/submissions.aspx>. (UNFCCC, 2015).

⁶ FOOTNOTE: UNFCCC. Greenhouse Gas Inventories, http://unfccc.int/national_reports/annex_i_ghg_inventories/national_inventories_submissions/items/8812.php. (UNFCCC, 2015).

⁷ FOOTNOTE: UNFCCC. National Communications Non-Annex 1, <http://unfccc.int/nationalreports/non-annexinatcom/submittednatcom/items/653.php> (UNFCCC, 2015).

⁸ FOOTNOTE: UNFCCC. National Communications Annex 1, <http://unfccc.int/nationalreports/annexinatcom/submittednatcom/items/7742.php>; (UNFCCC, 2015).

⁹ FOOTNOTE: UNFCCC. Biennial Update Reports, http://unfccc.int/national_reports/non-annex_i_natcom/reporting_on_climate_change/items/8722.php (UNFCCC, 2015).

1
2 Response options relying on the use of land could provide around a third of the additional mitigation needed
3 in the near term (2030) to close the gap between current policy trajectories based on NDCs and what is
4 required to achieve a 2°C (>66% chance) or 1.5°C (50 to 66% chance) pathway according to the UNEP
5 Emissions Gap Report (Roberts et al. 2006). The report estimates annual reduction potentials in 2030 from
6 agriculture 3.0 (2.3–3.7) GtCO₂-eq yr⁻¹, a combination of “uncertain measures” (biochar, peat-related
7 emission reductions, and demand-side management) 3.7 (2.6–4.8) GtCO₂-eq yr⁻¹; forests 5.3 (4.1–6.5)
8 GtCO₂-eq yr⁻¹, bioenergy 0.9 GtCO₂-eq yr⁻¹, and BECCS 0.3 (0.2 to 0.4) GtCO₂-eq yr⁻¹ (UNEP 2017) Table
9 4.1). These response options account for 35% of potential reduction (or 32% without bioenergy and BECCS)
10 out of a total (all sector) potential of 38 (35–41) GtCO₂-eq yr⁻¹. The potentials estimated in the UNEP
11 Emissions Gap Report are based on the technical potential of individual response options from literature
12 including that presented in Section 2.1. CDR related to land use, while not a substitute for strong action in
13 the energy sector, has the technical potential to balance unavoidable emissions that are difficult to eliminate
14 with current technologies (*high confidence*), with early action avoiding deeper and more rapid action later
15 (*very high confidence*) (Strefler et al. 2018; Elmar et al. 2018, SR15).
16
17
18

2.7 Plant and soil processes underlying land-climate interactions

Projecting future complex interactions between land and climate require Earth system models (ESMs). A growing number of studies suggested that many processes important for interactions between land and climate were missing in the CMIP5-class ESMs and that Dynamic vegetation models (DGVM) used tended to elevate CO₂ emission and removals (*high confidence*) (Busch and Sage 2017; Rogers et al. 2017; Anderegg et al. 2016; Tjoelker 2018; Sulman et al. 2014a; Wieder et al. 2018; Davidson et al. 2006a).

Ecosystem complexity stemming from the diversity of plants, animals and microbes, as well as their biological responses to gradual climate changes (e.g., adaptive migration) and disturbance events (e.g., extreme weather events, fire, pest outbreaks; Section 2.2), are of potential importance. Of these processes, this section focuses on plant and soil processes, as recent empirical work, including those explained in the following subsections, offer potential for improved model projections under warmer and CO₂ rich futures.

The magnitude of future uptake and release of CO₂ and other greenhouse gases by vegetation are among the greatest uncertainties (Ciais et al. 2013b). One reason for this uncertainty stems from the lack of understanding of the mechanisms responsible for plant responses to increasing temperatures. The short- and long-term projections of gross photosynthesis responses to changes in temperature, CO₂, nutrient availability vary greatly among the models (Busch and Sage 2017; Rogers et al. 2017). Net CO₂ exchange requires estimation of autotrophic respiration, which is another source of uncertainty in ESM projections (Malhi et al. 2011). The importance of plant acclimation of photosynthesis and respiration in understanding vegetation response to climate change is now widely recognised (*high confidence*) (Rogers et al., 2017; Tan et al., 2017; Tjoelker, 2018; Vanderwel et al., 2015 ; Section 2.7.1). Acclimation is broadly defined as the biochemical, physiological, morphological or developmental adjustments within the lifetime of organisms that result in improved performance at the new condition. Acclimation often operates over a time span of days to weeks, and can mitigate negative effects of climate change on organismal growth and ecosystem functions (Tjoelker 2018).

Soil carbon and microbial processes, which interact with plant responses to climate, represent another large source of uncertainty in model projections (*medium confidence*) (Sections 2.7.2, 2.7.3 and 2.7.4). Given the wide range of uncertainty associated with SOC size estimates, CMIP5 models use a wide range of starting SOC stocks from 510 to 3040 GtC (Todd-Brown et al. 2013). Soil microbial respiration is estimated to release 40–70 GtC annually from the soil to the atmosphere globally (Hawkes et al. 2017). Projections of changes in global SOC stocks during the 21st century by CMIP5 models also ranged widely, from a loss of 37 Gt to a gain of 146 Gt, with differences largely explained by initial SOC stocks, differing C input rates, and different decomposition rates and temperature sensitivities (Todd-Brown et al. 2013). With respect to land-climate interactions, the key processes affecting SOC stocks are warming (which is expected to accelerate SOC losses through microbial respiration) and acceleration of plant growth (which increases inputs of C to soils). However, complex mechanisms underlying SOC responses to moisture regimes, carbon addition, and warming drive considerable uncertainty in projections of future changes in SOC stocks (Sulman et al. 2014a; Singh et al. 2010; Wieder et al. 2018).

2.7.1 Temperature responses of plant and ecosystem production

Climate-change responses of net ecosystem production cannot be modelled by simple instantaneous response functions, because of thermal acclimation responses of plants and soil microbes, as well as delayed responses arising from interactions between plants and the soil (*high confidence*) (Slot et al. 2014; Rogers et al. 2017; Tan et al. 2017; Tjoelker 2018). Photosynthesis and respiration of component plant species exhibit different functional shapes among species (Slot et al. 2014), and carbon balance at the stand level is influenced by respiration of ecosystem biomass other than plants. Large uncertainty remains for thermal responses of bacteria and other soil organisms (Section 2.7.5). Bayesian statistical estimates of global photosynthesis and total ecosystem respirations suggest that they exhibit different responses to thermal anomaly during the last 35 years (Li et al. 2018b).

Thermal responses of plant respiration, which consumes approximately one half of GPP, have not been

1 appropriately incorporated in most ESMs (Davidson et al., 2006; Tjoelker, 2018). Assumptions associated
2 with respiration have been a major source of uncertainty for ESMs at the time of AR5. In most existing
3 models, a simple assumption that respiration doubles with each 10°C increase of temperature (i.e., $Q_{10} = 2$) is
4 adopted, ignoring acclimation. Even a small error stemming from this assumption can strongly influence
5 estimated net carbon balance at large spatial scales of ecosystems and biomes over the time period of
6 multiple decades (Smith and Dukes 2013; Smith et al. 2016b). In order to estimate more appropriate thermal
7 response curves of respiration, a global database including data from 899 plant species has been compiled
8 (Atkin et al. 2015), and respiration data from 231 plants species across seven biomes have been analysed
9 (Heskel et al. 2016). These empirical data on thermal responses of respiration demonstrate a globally
10 convergent pattern (Huntingford et al. 2017). According to a sensitivity analysis of a relatively small number
11 of ESMs, a newly derived function of instantaneous responses of plant respiration to temperature (instead of
12 a traditional exponential function of $Q_{10} = 2$) makes a significant difference in estimated autotrophic
13 respiration especially in cold biomes (Heskel et al. 2016).

14
15 Acclimation results in reduced sensitivity of plant respiration with rising temperature, i.e., down regulation
16 of warming-related increase in respiratory carbon emission (Atkin et al. 2015; Slot and Kitajima 2015;
17 Tjoelker 2018) (*high confidence*). For example, experimental data from a tropical forest canopy show that
18 temperature acclimation ameliorates the negative effects of rising temperature to leaf and plant carbon
19 balance (Slot et al. 2014). Analysis of CO₂ flux data to quantify optimal temperature of net primary
20 production of tropical forests also suggest acclimation potential for many tropical forests (Tan et al. 2017).
21 Comparisons of models with and without thermal acclimation of respiration show that acclimation can halve
22 the increase of plant respiration with projected temperature increase by the end of 21st century (Vanderwel et
23 al. 2015).

24
25 It is typical that acclimation response to warming results in increases of the optimum temperature for
26 photosynthesis and growth (Slot and Winter 2017; Yamori et al. 2014; Rogers et al. 2017). Although such
27 shift is a result of a complex interactions of biochemical, respiratory, and stomatal regulation (Lloyd and
28 Farquhar 2008), it can be approximated by a simple algorithm to address acclimation (Kattge et al. 2007).
29 Mercado et al., (2018), using this approach, found that inclusion of biogeographical variation in
30 photosynthetic temperature response was critically important for estimating future land surface carbon
31 uptake. In the tropics, CO₂ fertilisation effect (c.f., Box 2.3) is suggested to be more important for observed
32 increases in carbon sink strength than increased leaf area index or longer growing season (Zhu et al. 2016).
33 Acclimation responses of photosynthesis and growth to simultaneous changes of temperature and CO₂, as
34 well as stress responses above the optimal temperature for photosynthesis, remain a major knowledge gap in
35 modelling responses of plant productivity under future climate change (Rogers et al. 2017).

36 37 **2.7.2 Water transport through soil-plant-atmosphere continuum and drought mortality**

38
39 How climate change, especially changes of precipitation patterns, influence water transport through the soil-
40 plant-atmosphere continuum, is a key element in projecting the future of water vapour flux from land and
41 cooling via latent heat flux (Sellers et al. 1996; Bonan 2008a; Brodribb 2009; Choat et al. 2012; Sperry and
42 Love 2015; Novick et al. 2016; Sulman et al. 2016)(*high confidence*). Even without changes in leaf area per
43 unit area of land, when plants close stomata in response to water shortage, dry atmosphere, or soil moisture
44 deficit, the stand-level fluxes of water (and associated latent heat flux) decrease (Seneviratne et al. 2018).
45 Closing stomata enhances drought survival at the cost of reduced photosynthetic production, while not
46 closing stomata avoids loss of photosynthetic production at the cost of increased drought mortality (Sperry
47 and Love 2015). Hence, species-specific responses to drought, in terms of whether they close stomata or not,
48 have short and long-term consequences (Anderegg et al. 2018a; Buotte et al. 2019). Increased drought-
49 induced mortality of forest trees, often exacerbated by insect outbreak and fire (e.g., (Breshears et al. 2005;
50 Kurz et al. 2008a; Allen et al. 2010)) (Section 2.2.4), have long-term impact on hydrological interactions
51 between land and atmosphere (Anderegg et al. 2018b).

52
53 New models linking plant water transport with canopy gas exchange and energy fluxes are expected to
54 improved projections of climate change impacts on forests and land-atmosphere interactions (Bohrer et al.,
55 2005; Anderegg et al., 2016; Sperry and Love, 2015; Wolf et al., 2016)(*medium confidence*). Yet, there is
56 much uncertainty in the ability of current vegetation and land surface models to adequately capture tree

1 mortality and the response of forests to climate extremes like drought (Rogers et al. 2017; Hartmann et al.
2 2018). Most vegetation models use climate stress envelopes or vegetation carbon balance estimations to
3 project climate-driven mortality and loss of forests (McDowell et al. 2011); these may not adequately project
4 biome shifts and impacts of disturbance in future climates. For example, a suite of vegetation models was
5 compared to a field drought experiment in the Amazon on mature rainforest trees and all models performed
6 poorly in projecting the timing and magnitude of biomass loss due to drought (Powell et al. 2013). More
7 recently, the loss of water transport due to embolism (disruption of xylem water continuity) (Sperry and
8 Love 2015), rather than carbon starvation (Rowland et al. 2015), is receiving attention as a key physiological
9 process relevant for drought-induced tree mortality (Hartmann et al. 2018). A key challenge to modelling
10 effort is to consider differences among plant species and vegetation types in their drought responses. One
11 approach is to classify plant species to “functional types” that exhibit similar responses to environmental
12 variations (Anderegg et al. 2016). Certain traits of species, such as tree height, is shown to be predictive of
13 growth decline and mortality in response to drought (Xu et al. 2016a). Similarly, tree rooting depth is
14 positively related to mortality, contrary to expectation, during prolonged droughts in tropical dry forest
15 (Chitra-Tarak et al. 2017).

16 2.7.3 Soil microbial effects on soil nutrient dynamics and plant responses to elevated CO₂

17
18
19 Soil microbial processes influencing nutrient and carbon dynamics represent a large source of uncertainty in
20 projecting land-climate interactions. For example, ESMs incorporating nitrogen and phosphorus limitations
21 (but without considering the effects of mycorrhizae and rhizosphere priming) indicate that the simulated
22 future C-uptake on land is reduced significantly when both nitrogen and phosphorus are limited as compared
23 to only C-stimulation, by 63% (of 197 Pg C) under RCP2.6 and by 67% (of 425 Pg C) under RCP8.5 (Zhang
24 et al. 2013c). Mineral nutrient limitation progressively reduces the CO₂ fertilisation effects on plant growth
25 and productivity over time (Norby et al. 2010; Sardans et al. 2012; Reich and Hobbie 2013; Feng et al. 2015;
26 Terrer et al. 2017) (*robust evidence, medium agreement*). The rates at which nutrient limitation develops
27 differ among studies and sites. A recent meta-analysis shows that experimental CO₂ enrichment generally
28 results in lower nitrogen and phosphorus concentrations in plant tissues (Du et al. 2019a), and isotopic
29 analysis also suggest a global trend of decreases in leaf nutrient concentration (Craine et al. 2018; Jonard et
30 al. 2015). However, reduced responses to elevated CO₂ (eCO₂) may not be a simple function of nitrogen
31 dilution per se, as they result from complex interactions of ecosystem factors that influence nitrogen
32 acquisition by plants (Liang et al. 2016; Rutting 2017; Du et al. 2019a).

33
34 Increasing number of case studies suggest that soil microbial processes, such as nitrogen mineralisation rates
35 and symbiosis with plants, influence nutrient limitation on eCO₂ effects on plant growth (Drake et al. 2011;
36 Zak et al. 2011; Hungate et al. 2013; Talhelm et al. 2014; Du et al. 2019a) (*medium confidence*).
37 Rhizosphere priming effects (i.e., release of organic matters by roots to stimulate microbial activities) and
38 mycorrhizal associations are proposed to explain why some sites becoming nitrogen limited after a few years
39 and others sustaining growth through accelerated nitrogen uptake (Phillips et al. 2011; Terrer et al. 2017)
40 (*limited evidence, medium agreement*).

41
42 Model assessments that including rhizosphere priming effects and ectomycorrhizal symbiosis suggest that
43 soil organic matter (SOM) cycling is accelerated through microbial symbiosis (Elbert et al. 2012; Sulman et
44 al. 2017; Orwin et al. 2011; Baskaran et al. 2017) (*medium confidence*). Uncertainty exists in differences
45 among ectomycorrhizal fungal species in their ability to decompose SOM (Pellitier and Zak 2018) and the
46 capacity of ecosystems to sustain long-term growth with these positive symbiotic feedbacks is still under
47 debate (Terrer et al. 2017). ESMs include only biological nitrogen cycles, even though a recent study
48 suggests that bedrock weathering can be a significant source of nitrogen to plants (Houlton et al. 2018). In
49 contrast, rock weathering is widely considered to be the key for P availability, and tropical forests with
50 highly weathered soils are considered to be limited by P availability rather than nitrogen availability (Reed et
51 al. 2015). Yet, evidence from P-fertilisation experiments is lacking (Schulte-Uebbing and de Vries 2018) and
52 P limitation of tropical tree growth may be strongly species-specific (Ellsworth et al. 2017; Turner et al.
53 2018a). Limitation by availability of soil nutrients other than nitrogen and P has not been studied in the
54 context of land-climate interactions, except potassium (K) as a potentially limiting factor for terrestrial plant
55 productivity in interaction with N, P and hydrology (Sardans and Peñuelas 2015; Zhao et al. 2017; Wright et
56 al. 2018).

1
2 Anthropogenic alteration of global and regional nitrogen and P cycles, largely through use of chemical
3 fertilisers and pollution, has major implications for future ecosystem attributes, including C storage, in
4 natural and managed ecosystems (Peñuelas et al. 2013, 2017; Wang et al. 2017c; Schulte-Uebbing and de
5 Vries 2018; Yuan et al. 2018) (*high confidence*). During 1997-2013, the contribution of nitrogen deposition
6 to the global C sink has been estimated at $0.27 (\pm 0.13) \text{ GtC yr}^{-1}$, and the contribution of P deposition as
7 $0.054 (\pm 0.10) \text{ GtC yr}^{-1}$; these constitute about 9% and 2% of the total land C sink, respectively (Wang et al.
8 2017c). Anthropogenic deposition of nitrogen enhances carbon sequestration by vegetation (Schulte-
9 Uebbing and de Vries 2018), but this effect of nitrogen deposition on carbon sequestration may be offset by
10 increased emission of GHGs such as N_2O and CH_4 (Liu and Greaver 2009). Furthermore, nitrogen deposition
11 may lead to imbalance of nitrogen vs. phosphorus availability (Peñuelas et al. 2013), soil microbial activity
12 and SOM decomposition (Janssens et al. 2010), and reduced ecosystem stability (Chen et al. 2016b).

13 14 **2.7.4 Vertical distribution of soil organic carbon**

15
16 It has long been recognised that dynamics of soil organic carbon (SOC) represent a large source of
17 uncertainties on biogeochemical interactions of land with atmosphere and climate as detailed below. Since
18 AR5, there have been new understanding on SOC size, as well as microbial processes that influence SOM
19 dynamics under climate change and LULCC. Three existing data bases (SoilGrids, the Harmonized World
20 Soil Data Base, Northern Circumpolar Soil Database) substantially differ in estimated size of global soil
21 organic carbon (SOC) stock down to 1 m depth, varying between 2500 Pg to 3400 Pg with differences
22 among databases largely attributable to C stored in permafrost (Joosten 2015; Köchy et al. 2015; Tifafi et al.
23 2018). These values are four to eight times larger than the carbon stock associated with the terrestrial
24 vegetation (Bond-Lamberty et al. 2018). New estimates since AR5 show that much larger areas in Amazon
25 and Congo basins are peatlands (Gumbrecht et al. 2017; Dargie et al. 2019).

26
27 Deep soil layers can contain much more carbon than previously assumed (e.g., González-Jaramillo et al.,
28 2016) (*limited evidence, medium agreement*). Based on radiocarbon measurements, deep SOC can be very
29 old, with residence times up to several thousand years (Rumpel and Kögel-Knabner 2011) or even several
30 tens of thousands of years (Okuno and Nakamura 2003). Dynamics associated with such deeply buried
31 carbon remain poorly studied and ignored by the models, and not addressed in most of the studies assessed in
32 this subsection. Deep soil C is thought to be stabilised by mineral interactions, but recent experiments
33 suggest that CO_2 release from deep soils can also be increased by warming, with a 4°C warming enhancing
34 annual soil respiration by 34–37% (Hicks Pries et al. 2017) or addition of fresh carbon (Sebastien
35 Fontaine Pierre Barre, Nadia Bdioui, Bruno Mary, Cornelia Rumpel 2007). While erosion is not typically
36 modelled as a carbon flux in ESMs, erosion and burial of carbon-containing sediments is likely a significant
37 carbon transfer from land to ocean (Berhe et al. 2007; Asefaw et al. 2008; Wang et al. 2017e) (*medium
38 confidence*).

39 40 **2.7.5 Soil carbon responses to warming and changes in soil moisture**

41
42 Annually, 119 GtC is estimated to be emitted from the terrestrial ecosystem to the atmosphere, of which ca.
43 50% is attributed to soil microbial respiration (Auffret et al. 2016; Shao et al. 2013). It is yet not possible yet
44 to make mechanistic and quantitative projections about how multiple environmental factors influence soil
45 microbial respiration (Davidson et al. 2006a; Dungait et al. 2012). Soil warming experiments show
46 significant variability in temperature and moisture responses across biomes and climates; Crowther et al.,
47 (2016) found that warming-induced SOC loss is greater in regions with high initial carbon stocks, while an
48 analysis of an expanded version of the same dataset did not support this conclusion (Gestel et al. 2018).
49 Studies of SOC responses to warming over time have also shown complex responses. In a multi-decadal
50 warming experiment, Melillo et al., (2017) found that soil respiration response to warming went through
51 multiple phases of increasing and decreasing strength, which were related to changes in microbial
52 communities and available substrates over time. (Conant et al., (2011) and Knorr et al., (2005) suggested that
53 transient decomposition responses to warming could be explained by depletion of labile substrates, but that
54 long-term SOC losses could be amplified by high temperature sensitivity of slowly decomposing SOC
55 components. Overall, long-term SOC responses to warming remain uncertain (Davidson et al. 2006a;
56 Dungait et al. 2012; Nishina et al. 2014; Tian et al. 2015).

1
2 It is widely known that soil moisture plays an important role in SOM decomposition by influencing
3 microbial processes (e.g., Monard et al., 2012; Moyano et al., 2013; Yan et al., 2018), as confirmed by a
4 recent global meta-analysis (Hawkes et al. 2017) (*high confidence*). A likely mechanism involves that
5 increased soil moisture lowers C mineralisation rates under anaerobic conditions resulting in enhanced C
6 stocks, but experimental analyses have shown that this effect may last for only 3–4 weeks after which iron
7 reduction can actually accelerate loss of previously protected organic C by facilitating microbial access
8 (Huang and Hall 2017).
9

10 Experimental studies of responses of microbial respiration to warming have found variable results (Luo et al.
11 2001; Bradford et al. 2008; Zhou et al. 2011; Carey et al. 2016; Teramoto et al. 2016). No acclimation was
12 observed in C-rich calcareous temperate forest soils (Schindlbacher et al. 2015) and arctic soils (Hartley et
13 al. 2008), and a variety of ecosystems from the Arctic to the Amazon indicated that microbes appear to
14 enhance the temperature sensitivity of soil respiration in Arctic and boreal soils, thereby releasing even more
15 carbon than currently projected (Karhu et al. 2014). In tropical forests, P limitation of microbial processes is
16 a key factor influencing soil respiration (Camenzind et al. 2018). Temperature responses of symbiotic
17 mycorrhizae differ widely among host plant species, without a clear pattern that may allow generalisation
18 across plant species and vegetation types (Fahey et al. 2016).
19

20 Some new insights have been obtained since AR5 from investigations of improved mechanistic
21 understanding of factors that regulate temperature responses of soil microbial respiration. Carbon use
22 efficiency and soil nitrogen dynamics have large influence on SOC responses to warming (Allison et al.
23 2010; Frey et al. 2013; Wieder, William R., Bonan, Gordon B., Allison 2013; García-Palacios et al. 2015)
24 (*high confidence*). More complex community interactions including competitive and trophic interactions
25 could drive unexpected responses to SOC cycling to changes in temperature, moisture, and C inputs
26 (Crowther et al. 2015; Buchkowski et al. 2017). Competition for nitrogen among bacteria and fungi could
27 also suppress decomposition (Averill et al. 2014). Overall, the roles of soil microbial community and trophic
28 dynamics in global SOC cycling remain very uncertain.
29

30 **2.7.6 Soil carbon responses to changes in organic-matter inputs by plants**

31

32 While current ESM structures mean that increasing C inputs to soils drive corresponding increases in SOC
33 stocks, long-term carbon addition experiments have found contradictory SOC responses. Some litter addition
34 experiments have observed increased SOC accumulation (Lajtha et al. 2014b; Liu et al. 2009), while others
35 suggest insignificant SOC responses (Lajtha et al. 2014a; van Groenigen et al. 2014). Microbial dynamics
36 are believed to have an important role in driving complex responses to C additions. The addition of fresh
37 organic material can accelerate microbial growth and SOM decomposition via priming effects (Kuzyakov et
38 al. 2014; Cheng et al. 2017). SOM cycling is dominated by “hot spots” including the rhizosphere as well as
39 areas surrounding fresh detritus (*medium evidence, high agreement*) (Finzi et al. 2015; Kuzyakov and
40 Blagodatskaya 2015). This complicates projections of SOC responses to increasing plant productivity;
41 increasing C inputs could promote higher SOC storage, but these fresh C inputs could also deplete SOC
42 stocks by promoting faster decomposition (Hopkins et al. 2014; Guenet et al. 2018; Sulman et al. 2014b). A
43 meta-analysis by (van Groenigen et al. 2014) suggested that elevated CO₂ accelerated SOC turnover rates
44 across several biomes. These effects could be especially important in high-latitude regions where soils have
45 high organic matter content and plant productivity is increasing (Hartley et al. 2012), but have also been
46 observed in the tropics (Sayer et al. 2011).
47

48 Along with biological decomposition, another source of uncertainty in projecting responses of SOC to
49 climate change is stabilisation via interactions with mineral particles (Kögel-Knabner et al. 2008; Kleber et
50 al. 2011; Marschner et al. 2008; Schmidt 2011) (*high confidence*). Historically, conceptual models of SOC
51 cycling have centred on the role of chemical recalcitrance, the hypothesis that long-lived components of
52 SOC are formed from organic compounds that are inherently resistant to decomposition. Under the emerging
53 new paradigm, stable SOC is primarily formed by the bonding of microbially-processed organic material to
54 mineral particles, which limits the accessibility of organic material to microbial decomposers (Lützow et al.
55 2006; Keiluweit et al. 2015; Kallenbach et al. 2016; Kleber et al. 2011; Hopkins et al. 2014). SOC in soil
56 aggregates can be protected from microbial decomposition by being trapped in soil pores too small for

1 microbes to access (Blanco-Canqui and Lal 2004; Six et al. 2004) or by oxygen limitation (Keiluweit et al.
2 2016). Some new models are integrating these mineral protection processes into SOC cycling projections
3 (Wang et al. 2017a; Sulman et al. 2014b; Riley et al. 2014; Wieder et al. 2015), although the sensitivity of
4 mineral-associated organic matter to changes in temperature, moisture, fire (see Box 2.1) and carbon inputs
5 is highly uncertain. Improved quantitative understanding of soil ecosystem processes will be critically
6 important for projection of future land-climate feedback interactions.
7
8

9 **Frequently Asked Questions**

10 **FAQ 2.1 How does climate change affect land use and land cover?**

11
12
13 Contemporary land cover and land use is adapted to current climate variability within particular temperature
14 and/or rainfall ranges (referred to as climate envelopes). Anthropogenic greenhouse gas emissions impact
15 land through changes in the weather and climate and also through modifications in atmospheric composition
16 through increased greenhouse gasses, especially CO₂. A warming climate alters the current regional climate
17 variability and results in a shift of regional climate envelopes poleward and to higher elevations. The shift of
18 warmer climate envelopes into high latitude areas has potential benefits for agriculture here through
19 extended growing seasons and warmer seasonal temperatures and increased atmospheric CO₂ concentrations
20 enhances photosynthetic activity. However, this warming will also lead to enhanced snowmelt and reduced
21 albedo, permafrost melting and the further release of methane and CO₂ into the atmosphere as the permafrost
22 begins to decompose. Concurrent with these climate envelope shifts will be the emergence of new, hot
23 climates in the tropics and increases in the frequency, intensity and duration of extreme events (e.g. heat
24 waves, very heavy rainfall, drought). These emergent hot climates will negatively affect land use (through
25 changes in crop productivity, irrigation needs, management practices) and land cover through loss of
26 vegetation productivity in many parts of the world, and overwhelm any benefits to land use and land cover
27 derived from increased atmospheric CO₂ concentrations.
28

29 **FAQ 2.2 How do the land and land use contribute to climate change?**

30
31 Any changes to the land and how it is used can affect exchanges of water, energy, greenhouse gases (e.g.,
32 CO₂, CH₄, N₂O), non-greenhouse gases (e.g., biogenic volatile organic compounds – BVOCs), and aerosols
33 (mineral, e.g., dust, or, carbonaceous, e.g., black carbon) between the land and the atmosphere. Land and
34 land use change therefore alter the state (e.g., chemical composition and air quality, temperature and
35 humidity) and the dynamics (e.g., strength of horizontal and vertical winds) of the atmosphere, which in turn
36 can dampen or amplify local climate change. Land-induced changes in energy, moisture and wind can affect
37 neighbouring, and sometimes more distant, areas. For example, deforestation in Brazil warms the surface, in
38 addition to global warming, and enhances convection which increases the relative temperature difference
39 between the land and the ocean, boosting moisture advection from the ocean and thus rainfall further inland.
40 Vegetation absorbs carbon dioxide (CO₂) to use for growth and maintenance. Forests contain more carbon in
41 their biomass and soils than croplands and so a conversion of forest to cropland, for example, results in
42 emissions of CO₂ to the atmosphere, thereby enhancing the greenhouse gas-induced global warming.
43 Terrestrial ecosystems are both sources and sinks of chemical compounds such as nitrogen and ozone.
44 BVOCs contribute to forming tropospheric ozone and secondary aerosols, which respectively affect surface
45 warming and cloud formation. Semi-arid and arid regions release dust, as do cropland areas after harvest.
46 Increasing the amount of aerosols in the atmosphere impacts temperature in both positive and negative ways
47 depending on the particle size, altitude, and nature (carbonaceous or mineral for example). Although global
48 warming will impact the functioning and state of the land (see FAQ 2.1), this is not a one-way interaction as
49 changes in land and land use can also affect climate and thus modulate climate change. Understanding this
50 two-way interaction can help improve adaptation and mitigation strategies as well as manage landscapes.
51

52 **FAQ 2.3 How does climate change affect water resources?**

1
2 Renewable freshwater resources are essential for the survival of terrestrial and aquatic ecosystems and
3 human use in agriculture, industry and domestically. As increased water vapour concentrations are expected
4 in a warmer atmosphere, climate change will alter the hydrological cycle and therefore regional freshwater
5 resources. In general, wet regions are projected to get wetter and dry regions drier, although there are
6 regional exceptions to this. The consequent impacts vary regionally; where rainfall is projected to be lower
7 in the future (many arid subtropical regions and those with a Mediterranean climate), a reduction of water
8 resources is expected. Here increased temperatures and decreased rainfall will reduce surface and
9 groundwater resources, increase plant evapotranspiration and increase evaporation rates from open water
10 (rivers, lakes, wetlands) and water supply infrastructure (canals, reservoirs). In regions where rainfall is
11 projected to be higher in the future (many high latitude regions and the wet tropics), an increase in water
12 resources can be expected to benefit terrestrial and freshwater ecosystems, agriculture and domestic use,
13 however, these benefits may be limited due to increased temperatures. An increase in extreme rainfall events
14 is also expected which will lead to increases in surface runoff, regional flooding and nutrient removal as well
15 as a reduction in soil water and groundwater recharge in many places. Anthropogenic land use change may
16 amplify or moderate the climate change effect on water resources therefore informed land management
17 strategies need to be developed. A warming climate will exacerbate the existing pressures on renewable
18 freshwater resources in water-stressed regions of the Earth and result in increased competition for water
19 between human and natural systems.
20
21

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Appendix

This appendix provides all numbers that support **Figure 2.14** and **Figure 2.17** located in Section 2.5. It lists all model-based studies, with their references, that have been used to create the figures. Studies that examine the effects of historical and future scenarios of changes in anthropogenic land cover are presented in Table 1. The responses to idealised latitudinal deforestation and forestation can be found in Table A2.2.

The biophysical effects of changes in anthropogenic land cover reflect the impacts of changes in physical land surface characteristics such as albedo, evapotranspiration, and roughness length. The biogeochemical effects reflect changes in atmospheric CO₂ composition resulting from anthropogenic changes in land cover. The biogeochemical effects are estimated using three different methods:

1. Directly calculated within global climate models (Tables A2.1 and A2.2);
2. Calculated from off-line dynamic global vegetation models (DGVMs) estimates of net changes in the emissions of CO₂ from land (Table A2.1);
3. Calculated from observation-based estimates of net changes in the emissions of CO₂ from land (for historical reconstruction only, Table A2.1)

The mean annual and global temperature change (ΔT) resulting from biogeochemical effects is calculated as follows, for both DGVMs and observation-based estimates:

$$\Delta T = \Delta LCO_2 * TCRE$$

Where ΔLCO_2 is the cumulative changes in net emissions of CO₂ resulting from anthropogenic land cover changes during the time period considered (in Tera tons of carbon, TtC), and TCRE is the transient climate response to cumulative carbon emissions (Gillett et al. 2013; Matthews et al. 2009). TCRE is a measure of the global temperature response to cumulative emissions of CO₂ and has been identified as a useful and practical tool for evaluating CO₂-induced climate changes (expressed in °C per Tera tons of Carbon, °C/TtC). TCRE values have been estimated for a range of Earth system models (Gillett et al. 2013), (MacDougall et al. 2016). In the following, we use the 5th percentile, mean, and 95th percentile derived from the range of available TCRE values. For each DGVM or observation-based estimate, we then calculate three potential temperature change to bracket the range of climate sensitivities.

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Table A2.1: Model-based and observation-based estimates of the effects historical and future anthropogenic land cover changes have on mean annual global surface air temperature (°C). BGC and BPH correspond to the change in temperature resulting from respectively biogeochemical processes (e.g., changes in atmospheric CO₂ composition) and biophysical processes (e.g., changes in physical land surface characteristics such as albedo, evapotranspiration, and roughness length).

Reference of the study	Time period	Cumulative CO ₂ emissions from anthropogenic land cover change (TtC)	TCRE (°C/TtC)	Change in mean global annual (°C)	
				BGC	BPH
<i>Historical period (global climate models)</i>					
(Lawrence et al. 2018)	1850-2005	0.123	1.9	0.23	
(Simmons and Matthews 2016)	1750-2000 ¹⁰			0.22	-0.24
(Devaraju et al. 2016)	1850-2005	0.112	1.9	0.21	
(Zhang et al. 2013a)	1850-2005 ¹¹	0.097	1.75	0.17	-0.06
(Hua and Chen 2013)	~1850-2000 (average of two estimates)				-0.015
(Jones et al. 2013a)	Preindustrial (no exact dates)				-0.57
(Lawrence et al. 2012)	1850-2005	0.120	1.9	0.23	-0.10
(De Noblet-Ducoudré et al. 2012)	1972-2002 relative to 1900-1970				-0.042 ; -0.056 ; -0.005 ; -0.041 ; 0.021 ; -0.007 ; -0.005
(Pongratz et al. 2010)	20th century			0.16, 0.18	-0.03
(Arora and Boer 2010)	1850-2000	0.040,0.077	2.4	0.1, 0.18	
(Strengers et al. 2010)	20 th century				-0.06
(Kvalevåg et al. 2010)	Preindustrial (no exact dates)				+0.04 (CASE I)

¹⁰ FOOTNOTE: Land Use Change + Fossil Fuel emission simulation values are considered.

¹¹ FOOTNOTE: Carbon-Nitrogen-Phosphorous simulation values are considered.

(Findell et al. 2009)	1901-2004				+0.02
(Findell et al. 2007)	1990 relative to potential vegetation				+0.008
(Brovkin et al. 2006)	1700-1992 (5 models)				-0.24 ; -0.13 ; -0.14 ; -0.25 ; -0.17
(Betts et al. 2007; Betts 2001)	1750-1990				-0.02
(Hansen et al. 2005)	1880-1990				-0.04
(Feddema et al. 2005)	Preindustrial land-cover changes (no exact dates, "prehuman" simulations)				-0.39
(Matthews et al. 2004)	1700-2000 (average of 7 simulations)			0.3	-0.14
(Brovkin et al. 2004)	1800-2000			0.18	-0.26
(Zhao and Pitman 2002; Chase et al. 2000, 2001)	Preindustrial				+0.06
(Hansen et al. 1998)	Preindustrial land-cover changes				-0.14
Mean (\pmStandard deviation) of all studies				0.2\pm0.05	-0.1\pm0.14
<i>Historical period (Dynamic Global Vegetation Models/Bookkeeping model results)</i>					
(Li et al. 2017a)	1901-2012 (median of models)	0.148	0.88-1.72-2.52	0.13-0.25-0.37	
(Peng et al. 2017)	1850-1990 (realistic cases range)	0.087,0.139	0.88-1.72-2.52	0.1-0.15-0.22, 0.12-0.24-0.35	
(Arneeth et al. 2017a)	1901-2014 ¹²	0.089	0.88-1.72-2.52	0.1-0.15-0.22	
		0.210	0.88-1.72-2.52	0.18-0.36-0.53	
		0.179	0.88-	0.16-0.31-	

¹² FOOTNOTE: FLULCC,1 refers to land use change related fluxes accounting for new processes in their study.

			1.72-2.52	0.45	
		0.195	0.88-1.72-2.52	0.17-0.33-0.49	
		0.083	0.88-1.72-2.52	0.1-0.14-0.21	
		0.161	0.88-1.72-2.52	0.14-0.28-0.4	
		0.117	0.88-1.72-2.52	0.1-0.2-0.3	
		0.104	0.88-1.72-2.52	0.1-0.18-0.26	
		0.196	0.88-1.72-2.52	0.17-0.34-0.49	
(Pugh et al. 2015)	1850-2012 (gross land clearance flux)	0.157	0.88-1.72-2.52	0.14-0.27-0.39	
(Hansis et al. 2015)	1850-2012	0.269	0.88-1.72-2.52	0.19-0.36-0.53	
(Houghton et al. 2012a; Hansis et al. 2015)	1920-1999 (multi-model range)	0.072, 0.115	0.88-1.72-2.52	0.1-0.12-0.18, 0.1-0.2-0.3	
Mean (\pmStandard deviation) of all studies				0.24\pm0.12	
<i>Historical period (Observation-based estimates)</i>					
(Li et al. 2017a)	1901-2012	0.155	0.88-1.72-2.52	0.14-0.27-0.39	
(Li et al. 2017a; Avitabile et al. 2016; Carvalhais et al. 2014)	1901-2012 ¹³	0.160,0.165	0.88-1.72-2.52	0.14-0.27-0.40,0.14-0.28-0.41	
(Liu et al. 2015; Li et al. 2017a)	1901-2012	0.161,0.163	0.88-1.72-2.52	0.14-0.28-0.41	
(Le Quéré et al. 2015)	1870-2014	0.145	0.88-1.72-2.52	0.13-0.25-0.36	
(Carvalhais et al. 2014; Li et al. 2017a)	1901-2012	0.152,0.159	0.88-1.72-2.52	0.13-0.26-0.38, 0.14-0.27-0.4	
(Pan et al. 2011b; Li et al. 2017a)	1901-2012	0.119,0.122	0.88-1.72-2.52	0.10-0.20-0.30, 0.11-	

¹³ FOOTNOTE: Different harmonization methods. Method A assumes increase in cropland area in a grid cell taken from forest. Method C assumes increase in cropland and pasture taken from forest and then natural grassland if no more forest area available.

				0.21-0.31	
Mean (\pmStandard deviation) of all studies				0.25\pm0.10	
Future -RCP8.5 (global climate models)					
(Tharammal et al. 2018)	2006-2100	0.093	1.9	0.18	
(Lawrence et al. 2018)	2006-2100	0.211	1.9	0.40	
(Simmons and Matthews 2016)	2000-2100	-	-	0.35	-0.34
(Hua et al. 2015)	2006-2100	0.032	2.4	0.08	-
(Davies-Barnard et al. 2014)	2005-2100	0.02	2.1	0.04	-0.015
(Boysen et al. 2014; Quesada et al. 2017a; Brovkin et al. 2013b)	2005-2100	0.034	2.4	0.08	0.04
		0.025	2.1	0.05	0.0
		0.037	1.6	0.06	0.08
		0.062	2.2	0.13	-0.20
		0.205	1.6	0.33	-0.06
(Lawrence et al. 2012)	2006-2100	0.256	1.9	0.49	
Mean (\pmStandard deviation) of all studies				0.20\pm0.15	-0.1\pm0.14
Future -RCP8.5 (Dynamic Global Vegetation Model results)					
(Pugh et al. 2015)	2006-2100	0.169,0.171	0.88-1.72-2.52	0.15-0.29-0.42,0.15-0.29-0.43	
(IPCC 2014)	2005-2099	0.151	0.88-1.72-2.52	0.13-0.26-0.38	
Mean (\pmStandard deviation) of all studies				0.28\pm0.11	
Future RCP4.5 (global climate models)					
(Tharammal et al. 2018)	2005-2100	-0.029	1.9	-0.05	
(Lawrence et al. 2018)	2006-2100	0.053	1.9	0.10	
(Simmons and Matthews 2016)	2000-2100			0.37	-0.29
(Davies-Barnard et al. 2014)	2005-2100	-0.040	2.1	-0.08	0.14

(Lawrence et al. 2012)	2006-2100	0.148	1.9	0.28	
Mean (\pmStandard deviation) of all studies				0.12\pm0.17	-0.1\pm0.21
Future RCP4.5 (Dynamic Global Vegetation Model results)					
(Pugh et al. 2015)	2006-2100	0.016,-0.018	0.88-1.72-2.52	0.01-0.03-0.04,-0.02-(-0.03)-(-0.045)	
(IPCC 2014)	2005-2099	0.027	0.88-1.72-2.52	0.02-0.05-0.07	
Mean (\pmStandard deviation) of all studies				0.01\pm0.04	
Future RCP2.6 (global climate models)					
(Tharammal et al. 2018)	2005-2100	0.039	1.9	0.07	
(Simmons and Matthews 2016)	2000-2100			0.42	-0.35
(Hua et al. 2015)	2006-2100	0.036	2.4	0.09	
(Davies-Barnard et al. 2014)	2005-2100			0.04	-0.01
(Brovkin et al. 2013b)	2005-2100	0.039	2.4	0.09	
		0.019	2.1	0.04	
		0.065	2.2	0.14	
		0.175	1.6	0.28	
(Lawrence et al. 2012)	2006-2100	0.0154	1.9	0.03	
Mean (\pmStandard deviation) of all studies				0.13\pm0.12	-0.18\pm0.17
Future RCP2.6 (Dynamic Global Vegetation Model results)					
(Pugh et al. 2015)	2006-2100 (no harvest, managed cases)	0.057,0.084	0.88-1.72-2.52	0.05-0.09-0.14,0.07-0.14-0.21	
(IPCC 2014)	2005-2099	0.105	0.88-1.72-2.52	0.09-0.18-0.26	
Mean (\pmStandard deviation) of all studies				0.14\pm0.06	

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1 **Table A2.2: Model-based estimates of the effects idealised and latitudinal deforestation or forestation**
 2 **have on mean annual global and latitudinal surface air temperature (°C). BGC and BPH correspond**
 3 **to the change in temperature resulting from respectively biogeochemical processes (e.g. changes in**
 4 **atmospheric CO2 composition) and biophysical processes (e.g. changes in physical land surface**
 5 **characteristics such as albedo, evapotranspiration, and roughness length).**

Idealised Deforestation/Afforestation (global climate models)					
Reference	Change in forest area (Mkm ²)	Cumulative LCC flux (TtC)	TCRE (K/TtC)	Mean annual change in surface air temperature, averaged globally (and for the latitudinal band where trees are removed or added) (°C)	
				BGC	BPH
<i>Tropical Deforestation</i>					
(Devaraju et al. 2018b)	36.1				0.02 (1.14)
(Longobardi et al. 2016b)	23 ¹⁴	0.127	1.72	0.30	0.044 (-0.19)
(Devaraju et al. 2015c)	23			1.06	-0.04 (0.20)
(Brovkin et al. 2015)					-0.01,-0.13,-0.05
(Bathiany et al. 2010b)	23.1			0.40	0.18 (0.9)
(Snyder 2010)	23				0.2 (1.0)
(Bala et al. 2007)	23	0.418	1.72	0.72	0.70
(Voldoire 2006)					0.2,0.4,0.6
(Snyder et al. 2004b)	22.7				0.24(1.2)
(Claussen et al. 2001b)	7.5			0.19 (0.15)	-0.04 (0.13)
(Ganopolski et al. 2001c)	7.5				-0.5 (0.5)
(Henderson-Sellers and Gornitz 1984)					0.00
(Potter et al. 1981; POTTER et al. 1975)					-0.2
(Sagan et al.					-0.07

¹⁴ FOOTNOTE: For some studies that do not provide area deforested, IPSL-CM5 model grids used to calculate the area.

1979)					
Mean (\pm Standard deviation) of all studies				0.53 \pm 0.32	0.1 \pm 0.27(0.61 \pm 0.48)
Tropical Afforestation					
(Wang et al. 2014a) (Average of 4 simulations)					0.925
(Bathiany et al. 2010b)	23.1				-0.03 (-0.1)
Temperate Deforestation					
(Devaraju et al. 2018a)	18.8				0.18 (0.52)
(Longobardi et al. 2016b)	15	0.047	1.72	0.10	-0.077 (-0.22)
(Devaraju et al. 2015b)	15.3			0.39	-0.5 (-0.8)
(Bala et al. 2007)	15	0.231	1.72	0.40	-0.04
(Snyder et al. 2004b)	19.1				-0.22 (-1.1)
Mean (\pm Standard deviation) of all studies				0.29 \pm 0.13	-0.13 \pm 0.22 (-0.4 \pm 0.62)
Temperate Afforestation					
(Laguë and Swann 2016b)					0.3 (1.5)
(Wang et al. 2014a)					1.14
(Swann et al. 2012b)	15.3			-0.2, -0.7	0.3
(Gibbard et al. 2005)					0.27
Mean (\pm Standard deviation) of all studies				-0.45	0.50 \pm 0.36
Boreal Deforestation					
(Devaraju et al. 2018a)	23.5				-0.25 (-1.2)
(Longobardi et al. 2016b)	13.7	0.050	1.72	0.11	-0.38 (-0.9)

(Devaraju et al. 2015b)	13.7			0.06	-0.9 (-4)
(Dass et al. 2013)	18.5			0.12, 0.32	-0.35
(Bathiany et al. 2010b)	18.5	0.02	2.04	0.04	-0.28 (-1.1)
(Bala et al. 2007)	13.7	0.0105	1.72	0.02	-0.8
(Snyder et al. 2004b)	22.4				-0.77(-2.8)
(Caussen et al. 2001)	6			0.09 (0.12)	-0.23 (-0.82)
(Ganopolski et al. 2001c)	6				-1.0
Mean (\pm Standard deviation) of all studies				0.11 \pm 0.09	-0.55 \pm 0.29 (-1.8 \pm 1.2)
Boreal Afforestation					
(Bathiany et al. 2010b)					0.31 (1.2)

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