# Chapter 2: Land-Climate Interactions

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# Executive Summary

Land and climate interact in complex ways triggering a variety of feedbacks across multiple spatial and temporal scales. There is a range of coherence levels in the understanding of the land responses to climate change and its feedbacks to climate system in the recent studies and thus major uncertainty remains about climate-land interactions. Since AR5 a new understanding has emerged about ecological and hydrological processes shaping exchanges of water, energy, and greenhouse gases (GHGs) between land and atmosphere, which have potential to amplify or attenuate impacts of climate change on land or its feedbacks to climate.

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

**As a result of warming due to anthropogenic climate change, tropical and sub-tropical regions will see the emergence of novel climates beyond the envelope of current natural variability** *(medium evidence, high agreement)*. This, together with projected shifts in climate zones in mid and high latitudes, will expose regional ecosystems to disturbances beyond current regimes, which will alter the structure, composition and functioning of the systems. Additionally, in high-latitude areas warming is projected to accelerate permafrost thawing and to increase disturbance in boreal forests through abiotic (e.g., drought, fire) and biotic (e.g., pests, disease) agents {2.3.1, 2.3.2, 2.6.3}.

**Since the 1980s, global vegetation photosynthetic activity (i.e. greening) has increased primarily as a result of $CO_2$ fertilisation, nitrogen deposition, and climate change** (*medium evidence, high agreement*). However, since the mid-1990s, trends of decreased photosynthetic activity (browning) have increased, particularly in the northern hemisphere mid- to low-latitudes. Regional greening-to-browning reversals have been observed on all continents that have caused a slowdown in the global greening trend (*limited evidence, medium agreement*). Projections of global greening/browning trends are uncertain - projected regional increases of drought and heat waves in a warmer climate are linked to increased regional browning trends, however, in higher latitudes greening trends are projected to increase as a result of $CO_2$ fertilisation and longer growing periods {2.2, 2.3.3}.

**Climate change and increases in frequency, intensity and duration of extreme climate events are projected to substantially affect ecosystems and agricultural production** (*robust evidence, high agreement*). There is *high confidence* that heat waves will increase in frequency, intensity and duration into the 21$^{st}$ century in many regions of the world (Europe, North America, South America, Africa, Indonesia, the Middle East, south and south east Asia and Australia) and *medium confidence* that drought risk will increase over the Mediterranean region, central Europe, the Amazon and southern Africa. Combinatory drought/heat wave events, and the projected extended duration of these pose the greatest risk to terrestrial ecosystems in terms of gross primary productivity and results in a greater reduction in carbon sequestration (*medium evidence, medium agreement*) {2.3.5}.

**Wildfire regimes are being increasingly driven by changes in temperature** (*Medium evidence, medium agreement*), **in addition to droughts and human activity, with important implications for fires in a warming world.** Fire weather seasons have lengthened by 18.7% globally between 1979 and 2013. Although global land area burned may have declined in recent decades, mainly due to less burning in grasslands and savannahs, a surge in forest fires observed in recent years, especially in temperate and boreal North America, peatland fires, and drought plus land use change driven fires in tropical regions can result in net fire emissions of GHG and carbonaceous aerosols in the future {Cross-Chapter Box 3: Fire and Climate Change}.

***Biophysical and biogeochemical land forcing and feedbacks on climate system***

**Changes in regional climate that result from global warming, are dampened or enhanced by changes in local land cover and land use** (*medium evidence, medium agreement*). Decrease in maximum daytime surface temperature following irrigation for example can locally be as large as -3°C to -8°C while global warming induces smaller increases in this temperature indicator in some regions. Irrigation dampens warming during the growing season (*robust evidence, high agreement*). Deforestation in tropical regions will enhance GHG-induced surface warming in areas where trees have been removed (*robust evidence, high agreement*). Urbanisation will enhance warming in the city and its surroundings, especially during heat wave episodes (*robust evidence, high agreement*) {2.2, 2.6.1, 2.6.2}.

**Land surface processes modulate the likelihood, intensity and duration of many extreme events including heat waves, droughts** (*robust evidence, high agreement*) **and heavy precipitations** *(medium evidence, medium agreement).* Dry soil moisture anomalies favour summer heat wave conditions through reduced evapotranspiration (*robust evidence, high agreement*) and vegetation changes can amplify or dampen extreme events through changes in albedo and evapotranspiration (*medium evidence, medium agreement)* {2.6.1, 2.6.2, 2.6.3}.

**Land cover and land use changes affect both local and remote areas** *(robust evidence, medium agreement)*. These remote affects may be biogeochemical, for example as deforestation releases stored carbon into the atmosphere which is then globally distributed, or biophysical through (a) changes in convection that affect energy transfer upward and pole-ward and (b) changes in gradients of temperature, pressure and moisture which alter regional to hemispheric winds and consequently moisture and temperature advection at these scales**.** Spatial scales of these remote forcings range from neighbouring regions to hemispheric to global {2.4, 2.6.4}.

**Historical changes in land use caused significant changes in regional mean annual and seasonal surface air temperatures** *(robust evidence, high agreement*). **However, there is *no agreement* on whether it resulted in a discernible change of global mean annual surface air temperature** *(limited evidence, low agreement)*. The absence of global effect is the consequence of two opposing influences of land use on climate: net $CO_2$ source from land use and land cover change led to global warming *(robust evidence, high agreement)*, while increased land surface albedo in extra-tropical regions led to global annual cooling (*robust evidence, high agreement);* Land-use change induced reductions in evapotranspiration resulted in warming in many regions during the growing season *(medium evidence, medium agreement).* Under RCP8.5 scenario, model-based estimates do not indicate a major contribution from future land use changes to global annual surface air temperature increase, but indicate significant regional temperature increases *(limited evidence, medium agreement*) {2.6.1, 2.7}.

**Future climate-induced changes in land cover and functioning will have both positive and negative biophysical feedbacks on climate** *(robust evidence, high agreement).* Amplification or dampening of warming due to land feedbacks will differ by the region and season, and will depend on the concurrent changes in hydrological cycle *(robust evidence, high agreement).* In Arctic and Boreal regions the warming-induced greening, northward migration of tree line and thawing of snow and permafrost enhance winter, spring and summer warming (*robust evidence, high agreement*). In Tropical regions, where climate-induced reduction in rainfall are projected, the resulting land browning and reduction in tree cover will enhance warming and potentially further reduce rainfall (*limited evidence, medium agreement*) {2.2, 2.6.3}.

**While there was progress in quantifying regional emissions of anthropogenic and natural land aerosols (e.g. mineral dust; black, brown and organic carbon; biogenic volatile organic compounds), considerable uncertainty still remains about their historical trends, interannual and decadal variability, and about any changes in the future** (*medium evidence, medium agreement*). There are no direct observations of natural aerosols on global or regional scales. Emissions are derived either from remotely sensed observations of atmospheric concentrations of constituents or from top-down or bottom-up inventories or models. There is a growing recognition that regional climate is strongly affected by natural land aerosols *(medium evidence, medium agreement).* However, CMIP5-class models are limited in their abilities to represent land aerosols emissions, chemistry, and secondary aerosols production, and thus their feedbacks to climate {2.5}.

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

**Land is emitting and removing Greenhouse Gases (GHGs), which are affected simultaneously by natural and human drivers, making it difficult to separate anthropogenic fluxes of GHGs from natural** *(robust evidence, high agreement).* The emission and removals of $CO_2$, $N_2O$, and $CH_4$ are shaped by the effects of natural climate variability and natural disturbances, by the "indirect" anthropogenic effects of environmental change, and by the "direct" anthropogenic effects of Agriculture, Forestry and Other Land Use (AFOLU). AFOLU is a significant net source of GHG emissions *(robust evidence, high agreement)*, contributing around 24% *(medium evidence, medium agreement)* of anthropogenic emissions of $CO_2$, $CH_4$, and $N_2O$ combined. Estimating "anthropogenic" emission and removals of GHG is necessary in support of both the UNFCCC and the Paris Agreement. It is expected that the global stocktake will compare country reports of national Greenhouse Gas Inventories submitted to the UNFCCC with modelled mitigation pathways. This expectation implies a need to ensure consistency between, or reconciliation of, different approaches to estimating anthropogenic fluxes {2.4}.

**Global models estimate a net AFOLU emission of $CO_2$ of 4.9 ± 3.0 $GtCO_2$ $y^{-1}$ during 2007-2016 (about 12% of total anthropogenic $CO_2$ emissions)** *(robust evidence, medium agreement).* This global model flux based on bookkeeping models includes deforestation, afforestation and other land conversions as well as forest management and peatland fires, but it does not explicitly model the indirect anthropogenic effects of environmental change on managed land. The $CO_2$ net source is a combination of both gross emissions (e.g. due to forest harvest or burning for land clearing) and gross removals (e.g., due to afforestation and forest regrowth). The bookkeeping model-based AFOLU flux is broadly consistent with the mean of results from dynamic global vegetation models (DGVMs) that do include environmental change effects on managed land, but typically include fewer management processes {2.4}.

**Global vegetation models estimate a net removal of $CO_2$ of -11.2 ± 3.0 $GtCO_2$ $y^{-1}$ during 2007-2016 due to the indirect anthropogenic effects of global change on unmanaged lands (removing around 28% of total anthropogenic $CO_2$ emissions)** *(robust evidence, medium agreement).* The net effects of climate change, and the fertilising effects of rising atmospheric concentrations of CO2 and N (if included in the models) is removal of CO2 from the atmosphere. Thus when combining the direct and indirect effects on managed and unmanaged lands from DGVMs and bookkeeping models, during 2007–2016 the land was a sink of $CO_2$, -6.3 ± 3.0 $GtCO_2$ $y^{-1}$ *(robust evidence, medium agreement).* The modelled estimate of the total net global land-atmosphere flux is consistent with estimates based on atmospheric concentration measurements and inversion methods of -5.1 to -8.4 $GtCO_2$ $y^{-1}$ {2.4}.

**AFOLU global net $CO_2$ flux reported in national GHG Inventories was a source of 0.1 $GtCO_2$ $y^{-1}$ during 2005 to 2015, which is 4.7 $GtCO_2$ $y^{-1}$ lower compared to global bookkeeping modelling estimates over the same period, due primarily to conceptual differences in consideration of managed land fluxes.** All $CO_2$ emissions and removals on land defined by countries as managed land are considered anthropogenic, including those due to direct human activity and those due to environmental change. The discrepancy between global bookkeeping models' and GHG Inventories AFOLU flux is attributed to (1) GHG Inventories accounting for effects of both direct and indirect anthropogenic drivers on $CO_2$ flux (2) a far larger area of forests considered as "managed land" in GHG inventories than in models {2.4}.

**Land is a net source of $CH_4$, accounting for about 61% of anthropogenic $CH_4$ emissions during 2005 to 2015.** The major sources are natural wetlands (172–187 $TgCH_4$ $yr^{-1}$), and anthropogenic emissions from agriculture (137–140 $TgCH_4$ $yr^{-1}$), landfills (60 $TgCH_4$ $yr^{-1}$), and biomass burning (17 $TgCH_4$ $yr^{-1}$). AR5 attributed inter-annual variations in the atmospheric $CH_4$ accumulation rate to variability in natural wetland emissions, but new evidence points to the larger role of atmospheric processes, e.g. loss through reactions with OH radicals (*low confidence).* The soil $CH_4$ sink is increasing in response to increasing atmospheric

$CH_4$ concentrations (*low confidence*). Effects of changes in $CH_4$ sources due to changes in both temperature and rainfall are expected to play the key role from process understanding studies but are not well represented in the current models {2.4}.

**Agriculture is the main anthropogenic source of $N_2O$ due to fertiliser application and manure management (4.1 Tg $N_2O$-N $yr^{-1}$).** Emissions are largely from North America, Europe, East Asia, and South Asia, but emissions are growing across the tropics. Natural sources of $N_2O$ are estimated to be around 11 Tg $yr^{-1}$ and these sources have decreased by approximately 0.9 Tg $yr^{-1}$ due to tropical deforestation. Changes to $N_2O$ sources will likely be mainly a function of future use of N fertilisers and animal production systems (*medium confidence*). Climate change and its interaction with the development of more intensive agricultural systems are also important drivers of $N_2O$ flux, which is not estimated by models currently (*medium confidence*) {2.4}.

**Responses of vegetation and soil organic carbon (SOC) to rising atmospheric $CO_2$ concentration and climate change remain uncertain** *(robust evidence, high agreement*). Nutrient (e.g., Nitrogen, Phosphorus) availability can limit future plant growth and carbon storage under rising $CO_2$ (*robust evidence, high agreement*). However, new evidence suggests that ecosystem adaptation through plant-microbe symbioses could alleviate some nitrogen limitation (*medium evidence, high agreement*). SOC responses to changing climate, carbon input from enhanced plant growth, and human harvest is another key uncertainty for future GHG projections *(medium evidence, medium agreement*). Warming of soils and increased litter inputs will accelerate carbon losses through microbial respiration (*robust evidence, high agreement*). Thawing of high-latitude/altitude permafrost will increase rates of SOC loss and change the balance between $CO_2$ and $CH_4$ emissions (*moderate evidence, moderate agreement*). On the whole, projected climate change is expected to counteract potential benefits of $CO_2$ fertilisation {2.2}.

***Consequences for the climate system of land-based adaptation and mitigation options, including negative emissions***

**Response options on land could provide around a third of near-term 2030 to 2050 mitigation potential required to reach Paris Agreement targets based on assessments of the technical potential of individual options combined with integrated assessment model results** (*medium evidence, medium agreement)*. Assessments of the technical mitigation potential of individual response options are shown in Table ES.1. While some estimates include some sustainability and cost considerations, most do not include socio-economic barriers, the impacts of future climate change or non-GHG climate forcing. These are not all additive as some options compete for land and other resources, while others may reduce the demand for land {2.7.1, 2.7.3}.

**Table ES 2.1 Land-based mitigation options: reduction potential estimates**

| **Options** | **$GtCO_2e\ y^{-1}$** | **Confidence level** |
|---|---|---|
| Reduced deforestation | 1.0 - 5.8 | *robust evidence, high agreement* |
| Reduced forested degradation | 2.1 - 3.7 | *robust evidence, high agreement* |
| Afforestation/reforestation | 1.5 - 11.0 | *robust evidence, medium agreement* |
| Forest management | 2.0 - 5.8 | *robust evidence, medium agreement* |
| Peatland restoration | 0.8 - 3.2 | *low evidence, medium agreement* |
| Soil carbon sequestration in agriculture | 0.7 - 3.5 | *robust evidence, medium agreement* |
| Biochar | 1.0 - 2.6 | *medium evidence, medium agreement* |
| Bioenergy with and without carbon capture and storage | 0.5 - 12.0 | *medium evidence, low agreement* |
| Agricultural management | 1.5 - 2.0 | *robust evidence, medium agreement* |
| Reducing food and agricultural waste | 0.4 - 4.5 | *medium evidence, medium agreement* |
| Shifting to healthy diets | 2.2 - 6.4 | *robust evidence, high agreement* |

**Future Representative Concentration Pathway (RCPs) scenarios anticipate key contributions to climate change mitigation from land-based options, interlinked with other sectors** *(robust evidence, high agreement)*. Across a range of mitigation scenarios (RCP4.5, RCP2.6 and RCP1.9 respectively), most scenarios indicate strong reductions in $CO_2$ emissions due to avoided deforestation and carbon uptake due to afforestation. The level of carbon dioxide removal increases with the stringency of the climate target for both afforestation (1.3, 1.7 and 2.4 Gt $CO_2$ eq $yr^{-1}$ in 2100) and BECCS (6.5, 11 and 15.3 Gt $CO_2$ $yr^{-1}$ sequestered in 2100). $CH_4$ and $N_2O$ emissions are reduced compared to a no-mitigation baseline due to improved agricultural and livestock management: $CH_4$ emissions by 3.7, 3.0 and 2.1 Gt $CO_2$ eq $yr^{-1}$ in 2100; $N_2O$ emission are 2.0, 1.6 and 1.2 Gt $CO_2$ eq $yr^{-1}$ in 2100 – but emissions still persist until the end of the century. In addition, dietary shifts away from emission-intensive livestock products also lead to decreased $CH_4$ and $N_2O$ emissions. However, high levels of bioenergy production can result in increased $N_2O$ emissions due to N fertilisation of dedicated bioenergy crops {2.7.2}.

**Land-based mitigation in support of the Paris Targets will have large-scale consequences on the extent of forest cover and area under bioenergy crops, with implications for land carbon storage and biophysical effects on regional temperature** (*robust evidence, high agreement*). Climate change mitigation pathways can shape the land system dramatically as global forest area can change from about – 500 Mha up to + 1000 Mha in 2100 compared to 2010, and demand for 2nd generation bioenergy crops can range from less than 5000 up to about 20,000 million ton per year by 2100 in RCP2.6 scenarios, sourced from about 200–1500 Mha of land (*robust evidence; high agreement*). The net carbon effects of different options over time depend on where land change occurs and on prior land use. In high carbon lands such as forest and peatlands, the carbon benefits of land protection are greater in the short-term than converting land to bioenergy crops for BECCS, which can take several harvest cycles to “pay-back” the carbon lost (*medium evidence, medium agreement*). Afforestation/reforestation in tropical regions will dampen the regional warming by increasing evapotranspiration. While in the temperate to boreal regions, dampening would only occur during the growing season but additional regional warming would occur during the snowy season (predominantly due to decreased albedo) (*medium evidence, high agreement*) {2.6, 2.7.2}.

**Alternative model-based integrated pathways exist that achieve climate change targets with less need for land-demanding carbon dioxide removal (CDR).** Those rely on lifestyle changes and agricultural intensification in which reduced cattle stocks play an important role, with rapid and early reduction of GHG emissions and earlier CDR in the land but also in other sectors. (*robust evidence, high agreement*) {2.7.2}.

**About a quarter of the 2030 mitigation already pledged by countries under the Paris Agreement is expected to come from land-based mitigation measures** *(medium evidence, high agreement)*. Most of the Nationally Determined Contributions (NDCs) submitted by countries include land-based mitigation, mainly reduced deforestation and forest sinks. Few included soil carbon sequestration, agricultural management or bioenergy explicitly. Full implementation of country pledges (NDCs) is expected to result in net removal of 0.4 to 1.3 Gt$CO_2$ $y^{-1}$ in 2030 compared to the net flux in 2010 due to land-based mitigation (for pledges submitted up to February 2016, range represents low to high mitigation ambition in pledges, not uncertainty in estimates) {2.7.3}.

**Land sector carbon dioxide removal, while not a substitute for strong action in the energy sector, has the technical potential to balance unavoidable emissions, with early action avoiding deeper and more rapid action later** *(robust evidence, high agreement*). The Paris Agreement implies a need for anthropogenic removals to balance hard to eliminate emission (such as from air transport and food). There is sufficient technical potential for land sector carbon removals, but this would require strong early action to be a realisable potential on the short time-scales required, with potential consequences for land competition as well as other trade-offs, synergies and governance issues discussed elsewhere in the SRCCL {2.7}.

# 2.1 Introduction: Land – climate interactions

## 2.1.1 Climate determines land cover & land cover affects climate

**This chapter assesses the literature on two-way interactions between climate and land, with focus on scientific findings published since AR5** and aspects of the land-climate interactions that were not assessed in previous IPCC reports. Previous assessments mostly focused on the contribution of land to global climate change via its role in emitting or releasing greenhouse gases (GHGs) or via implications of changes in surface reflective properties (i.e., albedo) for radiative forcing. This chapter assess contributions of changes in land cover, use, and functioning for both global and regional climates. It examines science advances in our understanding of interactive changes of climate and land, including how climate change, variability and extremes influence managed and unmanaged lands and how changes land from direct (e.g., land use change and land management) and indirect (e.g., increasing atmospheric $CO_2$ concentration and nitrogen deposition) anthropogenic forcing affect climate system on local, regional, and global scale. In addition, Chapter 2 assesses knowledge gaps in land-climate interactions in modeling and observation.

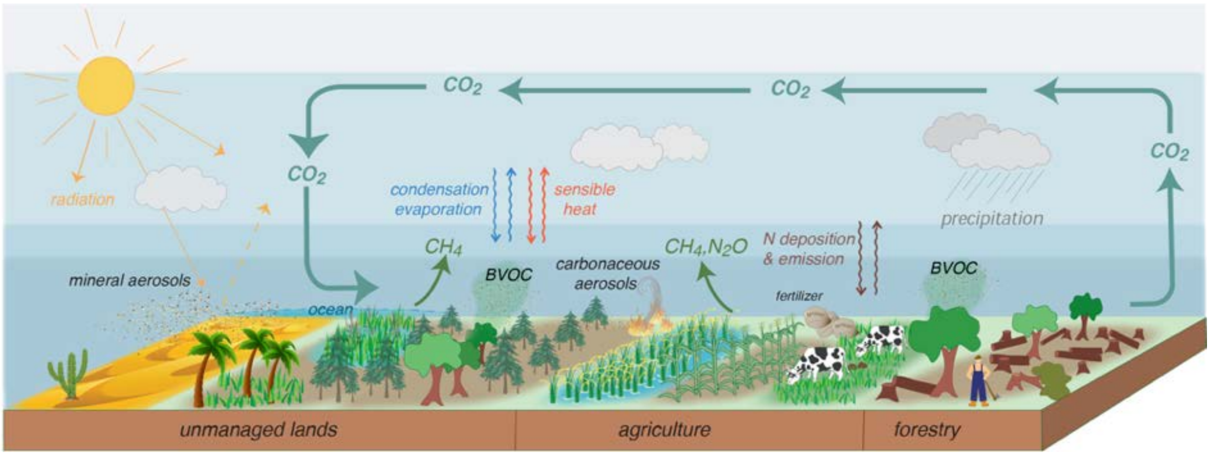


**Figure 2.1 The geographical distribution of different managed and unmanaged ecosystems, their physiological, ecological and hydrological state affect local, regional, and global climate. Land surface characteristics such as albedo and emissivity determine amount of short-wave and long-wave radiation from land to atmosphere. Surface roughness influences turbulent exchanges of momentum, energy, water and biogeochemical tracers. Land modulates the atmospheric composition, regional and global climate through emissions and removals of many GHGs, aerosols, and chemical tracers. Terrestrial aerosols affect both amount of surface radiation and precipitation through their role in clouds physics**

The chapter starts with a brief assessment of advances in understanding of processes underlying land-climate interactions, including those that may change from previous assessments (Section 2.2), followed by synthesis on the historical and projected responses of land patterns and functioning to climate change and extremes (Section 2.3). Subsequently, the chapter assess historical and future changes in terrestrial GHG (Section 2.4), non-GHG and aerosols (Section 2.5) exchanges between land and atmosphere on unmanaged and managed land. Section 2.6 focuses on how historical and future changes in land use and land cover influence climate change/variability through biophysical and biogeochemical forcing and feedbacks, how specific land management affect climate, and how in turn climate-induced land changes feedback on climate. Finally, Section 2.7 assesses consequences of land-based adaptation and mitigation options for the climate system in GHG and non-GHG exchanges and biophysical feedback. Sections 2.4 and 2.7 address implications of the Paris Agreement for land-climate interactions, and the scientific evidence base for ongoing negotiations around the Paris rulebook, the global stock take, and transparency and credibility in monitoring reporting and verification of the climate impacts of anthropogenic activities on land. It examines how land mitigation strategies may act on climate change through biophysical feedbacks and radiative forcing of land use changes on climate change and extremes (Sections 2.6), and concludes with policy relevant future changes in land use and sustainable land management for mitigation and adaptation (Section 2.7).

Since this is an assessment of land and climate *interactions*, the chapter also includes two boxes in order to integrate information across chapter sections. In this regard, boxes focus on processes, regions/biomes, and themes that are relevant to the climate-land interaction. These include boxes that focus on climate change and fire (Cross-Chapter Box 3: Fire and Climate Change) and on methodological approaches for estimating national to global scale anthropogenic land carbon fluxes (Box 2.1).

**Chapter 2 storylines**

**First, land and climate interact through a series of feedback loops**. Land and climate interact in a coupled way across spatial and temporal scales. Climate change/variability and extremes constrain terrestrial ecosystem functioning, land use/cover patterns, and land management options. Changes of land surface functioning and land use alter the land-atmosphere fluxes of GHGs/non-GHGs, water, and energy, and therefore, feedback to climate system. Biosphere–atmosphere feedbacks are considered as globally widespread, and explain up to 30% of precipitation and surface radiation variance in regions where feedbacks occur. Substantial biosphere–precipitation feedbacks are often found in regions that are transitional between energy and water limitation, such as semi-arid or monsoonal regions. Substantial biosphere–radiation feedbacks are often present in Mediterranean climate regions (Green et al. 2017).

**Second, terrestrial ecosystems and land use are affected by changing climate.** Global terrestrial ecosystems are sensitive to climate change/variability and extremes (Seddon et al. 2016). Climate change is expected to alter the distribution patterns of land cover (Schlaepfer et al. 2017), alter species composition and diversity, vegetation structure and productivity (Zhu et al. 2016), and nutrient and water cycles. The impacts of climate change on vegetation are reflected in a series of physiological processes, including changes in net plant carbon uptake, plant water use, plant growth and biomass allocation, competitive interactions, and responses to disturbances. However, data availability and science understanding on impacts of climate change on ecosystem and land use are highly heterogeneous across regions and biomes over the globe (Tica et al. 2016). Climate change is also reported to alter the seasonality of ecosystems at large scales (Gonsamo et al. 2017). Meanwhile, climate extremes are increasingly recognised as a driver behind interrupted changes of land surface through catastrophic disaster events (Lesk et al. 2016). Overall, climate change and extremes modify land productivity, the envelope of natural variability, phenology/seasonal distribution, and regime of disturbances and disasters. They put increased pressure on land management, compromise food security and ecosystem services, and intensity competition/conflict among land use, especially over ecotones and marginal lands.

**Third, land use/cover change and land management play an important and complex role in the climate system** (Pielke et al. 2016; Alkama and Cescatti 2016). They affect the climate via both biogeochemical and biophysical processes. Land is a source and a sink for several GHGs and aerosols and other non-GHG atmospheric constituents. Plus the nature of the land surface affects several biophysical properties and processes such as albedo, evapotranspiration, surface energy flux and alteration of energy partitioning of sensible and latent heat, and surface roughness (Burakowski et al. 2018), which in turn affect temperature, precipitation, humidity, cloud cover, and the planetary boundary layer at local, regional and global scales. Land surface processes also modulate the severity of heat waves (Wim et al. 2017), droughts, and other extreme events (Findell et al. 2017). Observed estimates of temperature change following deforestation indicate a smaller effect than model-based regional estimates in boreal regions, comparable results in the tropics, and contrasting results in temperate regions (Perugini et al. 2017). Recent satellite observation and model simulation suggest that Amazonian deforestation in the past three decades (Tyukavina et al. 2017) led to a shift towards a net carbon source (Baccini et al. 2017) and a dynamically driven hydroclimate (Pitman and Lorenz 2016; Zemp et al. 2017), with enhanced rainfall seen downwind of deforested areas (Lorenz et al. 2016a; Khanna et al. 2017). Similar impacts of deforestation are also found in west African rainforests (Klein et al. 2017). Satellite observations also reveal that the recent dynamics in global vegetation (Zhao et al. 2018) had contrasting biophysical impacts on the local climates, showing that the increasing trend in leaf area index (LAI) contributed to the warming of boreal zones through a reduction of surface albedo and to an evaporation-driven cooling in arid regions (Forzieri et al. 2017).

This chapter also pays special attention to advances in understanding scales, emerging issues, heterogeneity, and teleconnections.

**The biophysical impacts of land use change on climate are considered to be locally significant only (AR5), however, increasing evidence suggest that these impacts may go well beyond local level.** Changes of land use and land cover are reported at larger scales and extents than previously recorded, with recent advances in Earth observation and field network. Land cover change and land management can significantly affect surface energy and water balance through modification of albedo, evapotranspiration, surface roughness, and leaf area, and therefore, alter local and regional climate (De Vrese et al. 2016).

Meanwhile, **increasing evidence demonstrated the potential of sustainable land management in mitigating regional climate change** (Hirsch et al. 2017; Grassi et al. 2017). In the context of the Paris Climate Agreement, assuming full implementation of NDCs (Forsell et al. 2016), land use change could turn global land from a net anthropogenic source during 1990–2010 (1.3±1.1 $GtCO_2$-eq $yr^{-1}$) to a net anthropogenic sink of carbon by 2030 (up to-1.1-±-0.5 $GtCO_2$-eq $yr^{-1}$), and providing a quarter of emission reductions planned by countries (Grassi et al. 2017). However, negative emissions may be limited by biophysical and economic factors (Smith et al. 2016c).

**Major spatial heterogeneity exists**, for example multiple satellite-based analysis and modeling reveal complex climate effects of temperate forests and related energy budget (Ma et al. 2017b). Nevertheless, **major uncertainty still remains in our understanding of land-climate feedback** (Berg et al. 2017). State-of-art climate models (ESMs, GCMs, RCMs) show overall coherent biophysical behavior in response to land-use/cover changes: after tropical deforestation, increased albedo, reduction of evapotranspiration, decreased soil-moisture, increase in incoming radiation, enhanced surface and ground temperatures, wind-strengthening, less precipitation and clouds are robustly simulated (Pielke et al. 2011; Mahmood et al. 2014; Lawrence and Vandecar 2015; Lejeune et al. 2015, 2017; Quesada et al. 2017b; Devaraju et al. 2018).

### 2.1.2 Recap of previous IPCC and other relevant reports as baselines

Issues related to interactions between climate change and land surface processes in previous IPCC reports 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 various land use and cover, 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 Chapter 11. **Here, this chapter brings together land-related issues that cut across all three working groups, it also builds in previous special reports such as the Special Report on 1.5 degree warming (SR15), the Special Report on the ocean and cryosphere in a changing climate (SROCC), the Special Report on Renewable Energy and links to IPCC Guidelines on National Greenhouse Gas Inventories in the land sector. Meanwhile, this chapter goes further beyond that, as we assess knowledge that has never been reported in any of those previous reports.** We also try to reconcile the inconsistencies across the various IPCC reports, for example the glaring inconsistency between WGI and WGIII of AR5, whereas the LUC flux of WGI is interpreted in WGIII as the total AFOLU $CO_2$ balance.

**Here we briefly recapture key issues and findings from previous IPCC reports:**

<u>**GHGs and forcing**</u>: AR5 reported that atmospheric $CO_2$ and $CH_4$ increased by 40% from 278 ppm to 390.5 ppm and 150% from 722 ppb to 1803 ppb during 1750–2011, respectively. The $CO_2$ radiative forcing in AR5 (2011) is 1.82±0.19 W $m^{-2}$, an increase of 0.165 W $m^{-2}$ in relative to AR4 (2005) due to a 12ppm increases in atmospheric $CO_2$ mixing ratio. The $CH_4$ radiative forcing in AR5 is 0.48±0.5 W $m^{-2}$, an increase of 0.01W$m^{-2}$ in relative to AR4 due to 29 ppb increases in atmospheric $CH_4$ mixing ratio. **Annual net $CO_2$ emissions from anthropogenic land use change were 0.9 (0.1–1.7) GtC $yr^{-1}$ on average during 2002 to**

**2011 (*medium confidence*).** From 1750 to 2011, $CO_2$ emissions from fossil fuel combustion have released 375 [345–405] GtC to the atmosphere, while deforestation and other land use change are estimated to have released 180 [100–260] GtC. Of these cumulative anthropogenic $CO_2$ emissions, 240 (230–250) GtC have accumulated in the atmosphere, 155 (125–185) GtC have been taken up by the ocean and 160 (70–250) GtC have accumulated in terrestrial ecosystems (i.e., the cumulative residual land sink) (Ciais et al. 2013).

**<u>Terrestrial carbon source/sink</u>: Land carbon uptake projected among Climate Modelling Intercomparison Project Phase 5 (CMIP5) Earth System Models is very uncertain** due to the combined effects of climate change and land use change. There is *high confidence* that tropical ecosystems will uptake less carbon and there is *medium confidence* that at high latitudes, land carbon storage will increase in a warmer climate. Thawing permafrost in the high latitudes is potentially a large carbon source at warmer climate, but the magnitude of $CO_2$ and $CH_4$ emissions due to permafrost thawing is still uncertain. SR15 further indicates that constraining warming to 1.5°C would prevent the melting of an estimated permafrost area of 2 million $km^2$ over centuries compared to 2°C.

**<u>Land use change altered albedo</u>**: **AR5 provided *robust evidence* that anthropogenic land use change has increased the land surface albedo, which leads to an RF of $-0.15 \pm 0.10$ W $m^{-2}$**, however, it also indicated a large spread of estimates owing to different assumptions for the albedo of natural and managed surfaces and the fraction of land use changes before 1750. Generally, our understanding on albedo change due to land use alteration has enhanced from AR4 to AR5, with narrower range of estimates and higher confidence level. The radiative forcing of land use induced albedo change was estimated at -0.15 W $m^{-2}$ (-0.25 to about -0.05), with moderate confidence in AR5 (Myhre et al. 2013). While in AR4, the estimated radiative forcing was -0.2 W $m^{-2}$ (-0.4 to about 0), with *moderate-low confidence* (Forster, P., V. Ramaswamy, P. Artaxo, T. Berntsen, R. Betts, D.W. Fahey, J. Haywood, J. Lean, D.C. Lowe, G. Myhre, J. Nganga and G. Raga 2007).

**<u>Hydrologic feedback to climate</u>**: Land use change causes additional modifications that are not radiative, but impact the surface temperature, in particular through the hydrologic cycle. These are more uncertain and they are difficult to quantify, but tend to offset the impact of albedo changes. As a consequence, there is low agreement on the sign of the net change in global mean temperature as a result of land use change (Dentener et al. 2015).

**In terms of land-based water cycle changes, AR5 reported increased global evapotranspiration from the early 1980s to 2000s, however, the further increase is constrained due to lack of soil moisture availability.** The increasing aerosols level, declining surface wind speed and solar radiation are regionally dependent explanation to the decreasing evapotranspiration. In vegetated regions, rising $CO_2$ concentration can limit stomatal opening and thus transpiration as a main contribution to evapotranspiration. AR5 concluded increased global near surface air specific humidity since 1970. However, the moistening trend on land has abated since 2000, resulted in decreased near-surface relative humidity.

**<u>Climate-related extremes on land</u>: AR5 reported with *very high confidence* that impacts from recent climate-related extremes**, such as heat waves, droughts, floods, cyclones, and wildfires, **reveal significant vulnerability and exposure of some ecosystems to current climate variability**. Impacts of such climate-related extremes include alteration of ecosystems, disruption of food production and water supply, damage to infrastructure and settlements, morbidity and mortality, and consequences for mental health and human well-being. For countries at all levels of development, these impacts are consistent with a significant lack of preparedness for current climate variability in some sectors (Burkett et al. 2014). Recent SR15 report further indicates that limiting global warming to 1.5°C limits risks of increases in heavy precipitation events in several regions (*high confidence*).

**<u>Land-based *climate change mitigation*</u>*:* AR5 reported that adaptation and mitigation choices in the**

**near-term will affect the risks of climate change throughout the 21st century** (Burkett et al. 2014)**.** Agriculture, forestry and other land use (AFOLU) are responsible for about 10–12 $GtCO_2eq\ yr^{-1}$ anthropogenic greenhouse gas emissions mainly from deforestation and agricultural production. $CO_2$ emission from global forestry and other land use has declined since AR4, largely due to decreasing deforestation rates and increased afforestation. **Recent SR15 report** further assessed the role of land-based carbon-dioxide removal (CDR), and indicated that bioenergy with carbon capture and storage (BECCS) and afforestation are two CDR methods most often included in integrated pathways and that land use and land-use change emerge as a critical feature of virtually all mitigation pathways that seek to limit global warming to 1.5ºC.

Meanwhile, UNEP Global Environment Outlook (GEO-6) recently synthesised large-scale land surface changes, and concluded that the harvested crop area increased by 23% and global crop production rose by 87% between 1984 and 2015. In the 1990s, about 10.6 Mha $yr^{-1}$ of natural forests were lost. For the period 2010–2015, this rate had dropped to 6.5 Mha $yr^{-1}$. From 1975 to 2015, urban and settlements have expanded approximately 2.5 times, accounting for 7.6% of the global land area. Assessment based on satellite data shows that land degradation hotspots cover about 29 % of global land area (GEO-6 2018).

At a regional scale, 2018 Revision of World Urbanization Prospects indicates that 90% of future urban population growth (by 2050) will take place in Asia and Africa (UNDESA 2018). Meanwhile, wildland in Southeast Asia is deforested annually by more than 10,000 $km^2$, resulting in hundreds of millions of tons of carbon dioxide emissions per year between 2005 and 2015. 60% of the original mangroves in Southeast Asia has been cleared for coastal development (GEO-6 2018). From 2001 to 2013, cropland increased by 17% and pasture increased by 57% converted from forest in Latin America and the Caribbean. Deforestation to cropland from 1993 to 2013 is 405,000 ha in Canada, a much reduced deforestation rate compared to 1,286,000 ha from 1970 to 1990 (GEO-6 2018). In Africa, about 500,000 $km^2$ of land is degraded every year, with forest cover continually shrinking. The projected forest area is less than 6 million $km^2$ by 2050 due to the increasing conversion of forests to agricultural and housing need to support continuously increasing population (GEO-6 2018).

The way we manage our land is largely constrained by climate change and extremes, and also determine how effective we can adapt to climate change impacts and mitigate the warming risks. Our science knowledge in the processes has advanced for optimising our adaptation and mitigation efforts with coordinated and coherent land management cross sectors and stakeholder.

## 2.2 Progress in understanding of processes underlying land-climate interactions

AR5 concluded that ESMs were still at early stages in capturing processes shaping biogeochemical and biophysical interactions necessary for simulating the effects of land use, land use change and forestry (AR5, WG1, Section 6.3). More specifically, it listed six types of terrestrial biosphere processes that were yet to be adequately incorporated in models: (1) disturbances that influence resilience and recovery of forest carbon sinks, such as fire, logging harvesting, insect outbreaks and resulting variations in forest age structures, (2) decomposition processes in ecosystems with high organic carbon contents, including permafrost and wetlands (especially peatlands), (3) soil nutrient dynamics and their influence on vegetation functions, (4) impacts of tropospheric ozone and other pollutants, (5) coupling of water and heat transfer in soil-plant-atmosphere continuum, (6) surface transport of water and soil (via erosion). These processes interact with each other, and mostly their responses to climate factors are non-linear.

**Since AR5, a number of observational and modelling studies have refined understanding of how these processes could affect regional and global climate.** A growing number of studies suggest that diversity of plants, animals and microbes, ecosystem complexity and their biological responses, including adaptive migration and acclimation of organisms, play an important role and were missing in the AR5-class earth system models (ESMs). Furthermore, more studies have examined how human modifications of environment in rural and urban ecosystems affect land-climate interactions. This section provides an overview of the terrestrial processes that are receiving increasing attention or continue to be under intense scientific studies since AR5.

### 2.2.1 Biophysical and biogeochemical interactions

Climate and terrestrial biosphere interact with each other through heat exchanges and other physical interactions (biophysical interactions) and exchange of greenhouse gases and other chemicals between land and atmosphere (biogeochemical interactions). **'Biophysical interactions'** (Figure 2.2) depend on land surface characteristics such as reflectiveness of short-wave radiation (albedo) and surface roughness. There is *high confidence* that structure, density and seasonality of vegetation affect these physical interactions through their influences on albedo, latent heat, and turbulent and momentum fluxes. There is a *high agreement* that dense vegetation with high leaf area index (LAI), particularly forests, are prime regulators of the water cycle and heat transfer (Alkama & Cescatti, 2016; Ellison et al., 2017; Burakowski et al., 2018). Recent ESMs incorporate these biophysical interactions, demonstrating how effects of land use and land cover change might manifest beyond local scale (2.6).

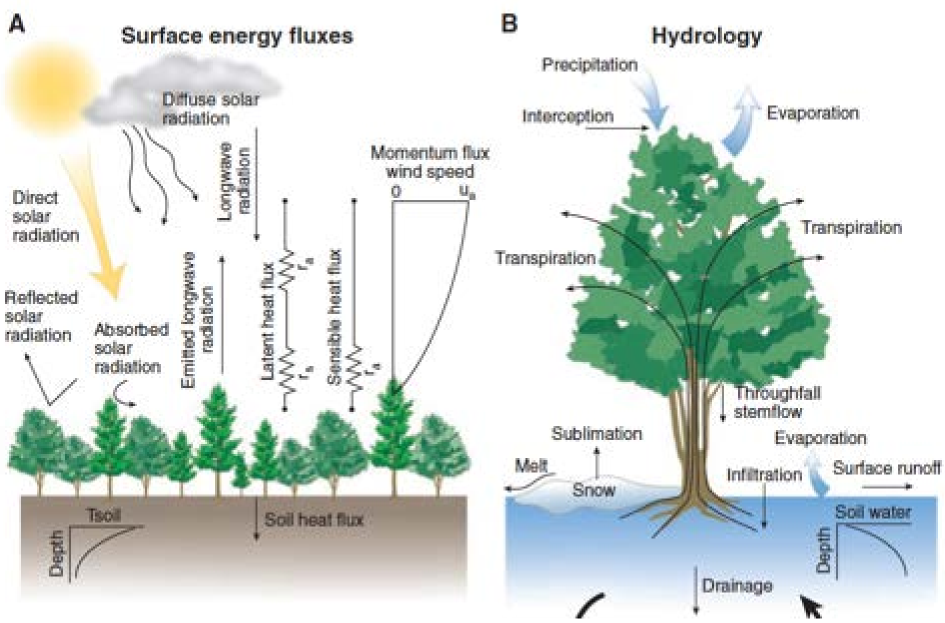


**Figure 2.2 Schematic of the biophysical exchanges that occur at the land (soil-vegetation) / atmosphere from (Bonan 2016). On the left components of the energy budget (radiation, turbulent and diffusive heat fluxes), on the right components of the water budget**

**''Biogeochemical interactions'** (2.1) encompass exchanges of greenhouse gases and aerosols between land and the atmosphere, which are determined by the structure and functioning of the terrestrial ecosystems (2.5). **Photosynthetic $CO_2$ uptake by forests has counteracted the atmospheric $CO_2$ increase by fossil fuel emission** (2.2.2, 2.4) (Los 2013; Erb et al. 2013; Zhu et al. 2016; Alton 2018; Schimel et al. 2015; Keenan and Riley 2018) (*robust evidence, high agreement*). However, future uptake and release of $CO_2$ and other greenhouse gases (e.g., $CH_4$ and $N_2O$) by vegetation are among the greatest uncertainties in order to adequately model the Earth's climate system (Ciais et al. 2013), as discussed in the remainder of this chapter. It is widely believed that since the 1960s, the land carbon sink has been increasing (Ballantyne et al. 2012) and reaching 3.1 ± 0.9 Pg C net removal of $CO_2$ from the atmosphere within 10 years, that is 25–30% of total anthropogenic emissions of carbon (2.1, 2.4). One reason for this uncertainty stems from the lack of understanding of the mechanisms responsible for enhanced carbon sink strength of land. Among potential mechanisms are enhanced vegetation growth under elevated atmospheric $CO_2$ and nitrogen (N) deposition (i.e., “fertilisation”)(2.2.2), re-growth of secondary forests recovering from prior harvesting and agricultural abandonment (Erb et al. 2013) (2.4), lengthening and warmer peak temperature of the growing season in the high latitudes (2.2.4).

Land use and land cover changes (LULCC) not only affect the atmospheric physical state and chemical composition locally, where such changes occur, but also remotely through cross-regional movements of atmosphere (i.e., atmospheric connections, see Section 2.6) and globally through their contribution to levels of greenhouse gases (Section 2.4). LULCC modulate the flux of fresh water, nutrients, and particulate matter from land to ocean, and consequently influence productivity and circulation patterns of the ocean (see subsections 2.5, 2.6). Conversion of forests to agriculture and infrastructure development (e.g., road construction and hydrological engineering) influence such phenomena (AR5, WGII). In the tropics, such conversion of forests to non-forests has resulted in large positive carbon emissions, particularly from conversion of peatland forests (van der Werf et al. 2010) (Section 2.4). Deforestation leads to soil erosion (Borrelli et al. 2017) and carbon loss (Jackson et al. 2017). Afforestation, reforestation and forest restoration reverse the flow of carbon and remove carbon from the atmosphere as forests grow (2.4), but there is much uncertainty involved (Cross-Chapter Box 1: Large Scale reforestation and afforestation, Chapter 1). Furthermore, fire suppression may lead to increased fuel load and wild fire in man-made forests, but such wildfires are difficult to incorporate into models (Cross-Chapter Box 3: Fire and Climate Change).

**There is *very high confidence* that** the magnitude and sign of the effects of LULCC differ among the regions. In tropical latitudes, deforestation causes decreases of evapotranspiration and latent heat transfer at local scales, and smoother land surface without trees reduces local convective rainfall (Khanna et al. 2017). In the temperate zones, deforestation has a cooling effect through reduced albedo and a warming effect through decreased evapotranspiration and latent-heat transfer, which offset each other (Findell et al. 2017). These complex interactions require improved understanding of the multiple factors involved in climate-land interactions across temporal and spatial scales.

**There is *very high confidence* that the magnitude and types of LULCC, as well as their interactions with climate change, differ among the regions** (Curtis et al. 2018; Song et al. 2018). In tropical latitudes, deforestation causes decreases of evapotranspiration and latent heat transfer at local scales, and smoother land surface without trees reduces local convective rainfall (Khanna et al. 2017). In the temperate zones, deforestation has a cooling effect through reduced albedo and a warming effect through decreased evapotranspiration and latent-heat transfer, which offset each other (Findell et al. 2017). These complex interactions require improved understanding of the multiple factors involved in climate-land interactions across temporal and spatial scales.

Certain types of changes in the land characteristics or GHG emissions, such as changes in the albedo or GHG emissions from the LULCC, are considered to be ‘forcings’ on the climate systems. Other changes, such as changes in the strength of land carbon sinks or sources are considered to be ‘feedbacks’ to the climate system – processes that modulate climate change response to ‘forcings’ (e.g., anthropogenic GHGs). An example of a negative biogeochemical feedback is enhanced uptake of atmospheric $CO_2$ by terrestrial biosphere (Ballantyne et al. 2012). An example of a positive biogeochemical feedback is a potential release of $CO_2$ and $CH_4$ from melting permafrost. In southernmost permafrost regions forest trees substantially delay thawing of permafrost and thus slow down impacts of climate warming (Baltzer et al. 2014). Coupled climate-biogeochemical ESMs are used to evaluate these multiple feedbacks between changing climate, vegetation and soils, and LULCC (Green et al. 2017). ESMs studies evaluate how different types of vegetation and soil processes such as photosynthesis, respiration, evapotranspiration, plant mortality or fires respond to changes in temperature, precipitation, short-wave radiation, etc. Variations among ESM results, such as prediction of atmospheric $CO_2$ are likely due to structural uncertainty of the land models (Hoffman et al. 2014).

## 2.2.2 Plant physiological responses and acclimations to increases in $CO_2$ and temperature

**There is *high agreement* that photosynthetic $CO_2$ exchange by plants is adequately approximated by the biochemical model of photosynthesis** developed by (Farquhar et al. 1980; Farquhar 1989), which incorporates the effects of atmospheric $CO_2$ concentration, water availability via stomatal adjustment (more in 2.2.5), and nutrient availability (more in 2.2.3) into of $CO_2$ uptake by plants (Farquhar 1989; Busch and Sage 2017). This model has been adopted in many climate-vegetation interaction models, and has allowed quantitative prediction of $CO_2$ fertilisation effects of increased photosynthetic production by terrestrial plants. Gross photosynthesis has less acute (but still significant) responses to temperature, compared to exponential increase of metabolic rate and respiration. At the time of AR5, short- and long-term responses of autotrophic respiration to warmer temperature regimes represented the largest source of uncertainty in estimation of Net Primary Production (NPP) in vegetation dynamic models (Masukume et al. 2017). However**, recent empirical work, including those explained in the following subsections, allows improved model prediction of photosynthesis-carbon balance in the warmer and $CO_2$ rich future as detailed in subsequent subsections.**

### *2.2.2.1 $CO_2$ fertilisation*

**There is *high confidence* that increasing atmospheric $CO_2$ could potentially enhance photosynthesis and water use efficiency of individual leaves, resulting in increased rates of plant growth and carbon sequestration** (FIELD et al. 1995; Swann et al. 2016). This is empirically supported by long term measurements of $CO_2$ and water vapour that show increases in net ecosystem carbon uptake and increases of photosynthetic water use efficiency in temperate and boreal forests (Keenan et al. 2013). A modelling study suggests that it may be possible for $CO_2$ fertilisation effects to ameliorate the impacts of future droughts and heat stresses in grassland net carbon uptake (Roy et al. 2016) .

**The realised $CO_2$ effect on growth observed in free-air carbon dioxide enrichment (FACE) experiments is highly variable due to nutrient limitations and other constraints on plant growth** (Körner, 2015; Feng et al. 2015; Paschalis et al. 2017; Terrer et al. 2017; Du et al. 2019). While elevated $CO_2$ does indeed result in increased short-term $CO_2$ uptake per unit leaf area, it is not certain whether this translates to increased growth rates of the whole plant, vegetation communities, ecosystems and biomes at decadal or longer time scales. Apparent inconsistencies among studies stem from multiple factors that constrain plant growth, such as whole-plant resource allocation constraints, nutrient limitation (especially N, P and possibly K), plant and soil water balance, light limitation, soil organic matter decomposition, and changes in plant community composition, mortality rates and biomass turnover (Körner 2006, Penuelas et al. 2017). Some species capable of overcoming N limitation by mutualistic association between roots and microbes can overcome nutrient limitation on $CO_2$ fertilisation effect (Terrer et al. 2017). Plant root-soil-microbe interaction plays an important role in mediating the nature of nutrient acquisition for supporting enhanced productivity under increasing $CO_2$ levels (2.2.3).

**There is *high confidence* that a consistent consequence of elevated $CO_2$ is increased plant water-use efficiency through optimal stomatal regulation**, possibly enhancing drought tolerance of plants (Berry et al. 2010; Ainsworth and Rogers 2007). Long-term $CO_2$ and water vapour flux measurements show that water-use efficiency in temperate and boreal forests of the Northern Hemisphere has increased over the past two decades more than predicted by the theory and terrestrial biosphere models (*high agreement, robust evidence*) (Keenan et al. 2013; Laguë and Swann 2016). Further, this increase has been accompanied by enhanced photosynthetic production and decreased evapotranspiration at the ecosystem level. Studies use tree ring records to pick up possible signals of the $CO_2$ fertilisation effect over time scales of several decades to a century or more. Increased water-use efficiency (of 30–35% over the past 150 years) is inferred from stable carbon isotope ratios in tree rings in three tropical wet forests across the globe, but there is no evidence for any acceleration of tree growth rates in support of the $CO_2$ fertilisation effects (van der Sleen et al. 2014). Experimental work at three FACE sites in the temperate region, in which tree ring and a combination of stable carbon and oxygen isotopes of wood are analysed, confirms increased water use efficiency over time (Battipaglia et al. 2013). There is, however, considerable uncertainty whether such increase in leaf-level water use efficiency actually translates into plant growth and benefits in NPP across ecosystems and biomes. At the levels of whole plant and vegetation, growth is constrained by availability of soil nutrients, in particular nitrogen and phosphorus (Reich and Hobbie 2013; Norby et al. 2010) (Section 2.2.3).

**There is *high agreement* that regional differences should be considered in order to evaluate the total effects of warming, anthropogenic $CO_2$ fertilisation, precipitation change, and nitrogen deposition**. Newer statistical treatment of tree ring and isotope data from 37 published studies representing all the major biomes concluded that intrinsic water use efficiency (iWUE) was consistently increasing (in the range of 10–60% during 1960–2010) but this did not necessarily translate into enhanced tree growth rates (Silva and Anand 2013). Only in boreal, alpine and Mediterranean trees about an increase of 20% in iWUE resulted in increased growth; indeed, in other major biomes including temperate, subtropical and tropical forests, there was actually a negative relationship between iWUE and tree growth rate. Analyses of records from the International Tree Ring Data Bank (ITRDB) indicated that only about 20% of the global sites showed increasing trends in tree growth that cannot be explained by climate variability, N deposition, elevation or latitude; thus this is taken as evidence for direct $CO_2$ fertilisation of forests on a limited scale during the 20th century (Gedalof and Berg 2010). Local site conditions and individual species' responses would determine future forest dynamics and the nature of the $CO_2$ effect.

**Approaches using top-down analyses from satellite data of changes in global vegetation greenness and global mass balance model argue that significant increases in NPP can be attributed to $CO_2$ fertilisation** (Alton 2018; Keenan and Riley 2018) (*high agreement*, *robust evidence*). A global analysis using a "reconstructed vegetation index" (RVI) for the period 1901–2006 from MODIS satellite-derived NDVI (Normalized Vegetation Difference Index) suggests that $CO_2$ fertilisation contributed at least 40% of the observed increase in land carbon sink strength (Los 2013). Intercomparison of ESMs suggests that 60% of the recent terrestrial carbon sink can be directly attributed to increasing atmospheric $CO_2$ (Schimel et al. 2015). Models also suggest that relative importance of $CO_2$ fertilisation is greater in the tropics, compared to higher latitudes where increased greenness from warmer climate and longer growing season are the main causes of observed increases in vegetation productivity (Zhu et al. 2016). Improved detection and attribution methods, in combination with higher-resolution remotely sensed data, are reducing the discrepancies between bottom-up estimates from ground-based measurements of vegetation productivity versus top-down approaches (Saeki and Patra 2017) (see Section 2.4).

#### *2.2.2.2 Acclimation and other physiological responses*

Plants can acclimate to changing $CO_2$ levels, temperature, and other environmental conditions through adjustment of gene expressions and physiological mechanisms. 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). Optimal temperature for forest NPP (Tan et al. 2017) is determined as the combined result of temperature responses of photosynthesis and respiration, which exhibit different functional shapes. Additional source of uncertainty comes from the fact that 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 2.2.6). 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 respiration and photosynthesis have not been appropriately incorporated in most ESMs** (*high agreement*). Assumptions associated with respiration have been a major source of uncertainty for ESMs at the time of AR5. In most existing models, a simplistic assumption that respiration doubles with each 10°C increase of temperature (i.e., $Q_{10} = 2$) is adopted, ignoring acclimation. Such assumption on thermal responses of respiration can strongly influence estimated net carbon balance at large spatial scales of ecosystems and biomes, as well as over the time period of multiple decades (Smith and Dukes 2013; Smith et al. 2016b). To amend this situation, global database (GlobResp) has been compiled to amend such data deficiency, leading to meta-analysis of 899 plant species, spanning a range of plant functional types and biomes from sea level to 3450m above sea level (Atkin et al. 2015), and another of 231 plants species across seven biomes (Heskel et al. 2016). The empirical data generated during the last decade on thermal responses of respiration demonstrate a globally convergent empirical pattern (Huntingford et al. 2017). According to a sensitivity analysis, a newly derived function of instantaneous responses of plant respiration to temperature (instead of a traditional exponential function of $Q_{10} = 2$) makes a significant reduction of autotrophic respiration especially in cold biomes (Heskel et al. 2016).

**Acclimation results in reduced sensitivity of respiration with rising temperature, i.e., down regulation of warming-related increase in respiratory carbon emission in all biomes** (Slot and Kitajima 2015; Tjoelker 2018) (*high agreement, robust evidence*). For example, experimental data from a tropical forest canopy show that temperature acclimation ameliorates the negative effects of rising temperature to leaf and plant carbon balance (Slot et al. 2014). Comparisons of models with and without thermal acclimation of respiration show that acclimation can halve the increases of plant respiration with predicted temperature increase by the end of 21st century (Vanderwel et al. 2015).

**In response to warming, there is *high agreement* that the optimum temperature for photosynthesis shifts towards the acclimation temperature, but across species and functional groups, optimum temperature (Topt) for photosynthesis generally does not increase as much as growth temperature** (Slot and Winter 2017; Yamori et al. 2014). The shift in Topt is underpinned by a complex interaction of biochemical, respiratory, and stomatal regulation (Lloyd and Farquhar 2008), but (Kattge and Knorr 2007) provide a simple algorithm to address acclimation. (Mercado et al. 2018), using this approach, found that inclusion of biogeographical variation in photosynthetic temperature response was critically important for estimating future land surface carbon uptake. This stresses the need for empirical data on acclimation of photosynthesis from across the globe, but especially from tropical forests, which are currently lacking. This is a gap in knowledge, given that in the tropics $CO_2$ fertilisation effects is suggested to be more important for observed increases in carbon sink strength than increased greening *per se* productivity (Zhu et al. 2016). Acclimation to simultaneous changes of temperature and $CO_2$ are even less well understood and represented in the models.

### 2.2.3 Nutrient limitation of plant growth

**Limitation in mineral nutrient availability, especially of soil nitrogen, is a factor that may reduce the $CO_2$ fertilisation effects plant growth and productivity over time** (Feng et al. 2015; Terrer et al. 2017) (*robust evidence, high agreement*). Yet, the mechanisms and factors leading nitrogen limitation may differ among plant species and ecosystems. The stoichiometry of C:N:P would eventually determine the upper limit of growth responses of individual plants and ecosystem carbon sequestration to increasing $CO_2$ (Sardans et al. 2012). A recent meta-analysis supports this perspective that experimental $CO_2$ enrichment generally results in lower N and P concentrations in plant tissues (Du et al. 2019). However, reduced responses to elevated $CO_2$ ($eCO_2$) may not be a simple function of N dilution *per se*, but instead they result from complex interactions of ecosystem factors that influence N acquisition by plants (Liang et al. 2016; Rutting 2017).

Experimental evidence for progressive N limitation of $eCO_2$ responses exists for both forest (Norby et al. 2010) and grassland (Reich and Hobbie 2013) ecosystems (*medium evidence, medium agreement*). Yet, there is considerable uncertainty in long-term responses of various ecosystems to interactive effects of $eCO_2$ and nutrient availability. At the Oak Ridge FACE experiment site in a temperate deciduous forest, NPP significantly increased under exposure to 550 ppm relative to the ambient control by 24% during 2001–2003 but then declined to only 9% by 2008; this could be explained by declining N availability (Norby et al. 2010). Similarly, lowered soil N availability in a long-term temperate grassland experiment halved $CO_2$ fertilisation effect during 2001–2010 compared to the previous decade under $eCO_2$ (Reich and Hobbie 2013). $CO_2$ fertilisation effect is expected to be proportionally large in semi-arid habitats where plant water-efficiency because $eCO_2$ reduces stomatal limitation on photosynthesis (including C4 grasses that are not expected to benefit much from $eCO_2$)(Donohue et al. 2013; Morgan et al. 2011; Derner et al. 2003). But, under N limitation, effects of $eCO_2$ were unobserved in dry summers (Reich et al. 2014). After more than 12 years of $eCO_2$ growth of C4 grasses unexpectedly started to be stimulated, while growth of C3 grasses was the same as at ambient $CO_2$, because N mineralisation increased in C4 grass plots but not in C3 grass (Reich et al. 2018). Hence, much uncertainty remains about interactive effects of $eCO_2$, N limitation, moisture availability, temperature, and plant functional types on carbon sequestration in future.

**Soil microbial processes, such as nitrogen mineralisation rates and symbionts with plants, influence N limitation on $eCO_2$ effects on plant growth and stoichiometry** (Du et al. 2019) (*medium evidence, medium agreement*). In the Duke FACE experiment, accelerated soil N cycling supported increased N uptake and growth enhancement (Drake et al. 2011). Similar results were observed in an aspen forest (Talhelm et al. 2014; Zak et al. 2011) and in an oak woodland (Hungate et al. 2013). This apparent contradiction (with some sites becoming nitrogen limited after a few years and others sustaining growth through accelerated N uptake) is likely be explained by rhizosphere priming effects and mycorrhizal associations (Terrer et al. 2017) (*limited evidence*). Measurements of root exudation suggest an alternative explanation that enhanced root exudation under elevated $CO_2$ can accelerate soil N cycling (Phillips et al. 2011).

**Model assessments that included rhizosphere priming effects and ectomycorrhizal symbioses suggest that acceleration of soil organic matter (SOM) cycling through microbial symbiosis could explain enhanced N availability and plant growth** (Sulman et al. 2017; Orwin et al. 2011; Baskaran et al. 2017) (*medium evidence, medium confidence*) (2.2.6). Uncertainty exists in differences among ectomycorrhizal fungal species in their ability to decompose SOM (Pellitier and Zak 2018) and the capacity of ecosystems to sustain long-term growth with these positive symbiotic feedbacks is still under debate (Terrer et al. 2017). Most N cycle models focus on biological factors, including the effects of symbiotic N fixation including those found in rock crust and bryophyte and lichen carpets (Elbert et al. 2012). A recent study by Houlton et al. (2018) suggests that bedrock weathering is a significant source of nitrogen to plants, accounting for 19 to 31 Tg $yr^{-1}$ of nitrogen mobilisation, but existing ESMs include only biological N cycles. It is widely accepted that input of phosphorus to the soil largely comes from rock weathering, and P differs from N in how it limits plant productivity and carbon cycles (Reed et al. 2015). Compared to boreal and temperate ecosystems, it is often presumed that tropical forests with highly weathered soils are limited by P availability rather than N (reviewed by Reed et al. 2015), but evidence from P-fertilisation experiments is lacking (Schulte-Uebbing and de Vries 2018) and P limitation of tropical tree growth is strongly species-specific (Turner et al. 2018). Zhang et al. (2013) reports that ESMs incorporating N and P limitations indicate that the simulated future C-uptake on land was reduced significantly when both N and P are limited as compared to only C-stimulation, by 63% (of 197 Pg C) under RCP2.6 and by 67% (of 425 Pg C) under RCP8.5.

**There is *high confidence* that anthropogenic alteration of global and regional N and P cycles, largely through use of chemical fertilisers and pollution, has major implications for future C storage in natural and managed ecosystems** (Peñuelas et al. 2013; Peñuelas et al. 2017; Wang et al. 2017; Schulte-Uebbing and de Vries 2018; Yuan et al. 2018) *(robust evidence, high agreement).* During 1997-2013, the contribution of N deposition to the global C sink has been estimated at 0.27 (± 0.13) Pg C $yr^{-1}$, and the contribution of P deposition as 0.054 (± 0.10) Pg C/ $yr^{-1}$; these constitute about 9% and 2% of the total land C sink, respectively (Wang et al. 2017c). There is *robust evidence* for anthropogenic deposition of N depositions enhancing carbon sequestration by vegetation (Schulte-Uebbing and de Vries 2018). However, this effect of N depositions on carbon sequestration may offset by increased emission of GHGs such as $N_2O$

and $CH_4$ (Liu and Greaver 2009). Furthermore, N depositions may lead to imbalance of nitrogen *vs.* phosphorus availability (Peñuelas et al. 2013) and reduced ecosystem stability (Chen et al. 2016c). Limitation by availability of soil nutrients other than N and P has not been studied as much. But, a strong case has also been recently made for potassium (K) as a possible limiting factor for plant productivity in terrestrial ecosystems especially in water-limited systems; N deposition has inhibitory effects on K availability, and thus interactive effects of N, P, K and water need to be incorporated into ESMs (Sardans and Peñuelas 2015).

### 2.2.4 Seasonality of ecosystem processes relevant for land-atmosphere interactions

**There is *robust evidence and high agreement* that phenology, that is seasonal activities of organisms, respond to environmental cues such as temporal patterns of temperature, day length, and moisture.; shifts in phenological timing have been documented in various organisms in the past half century in response to on-going climate change** (Peñuelas et al. 2002; Gordo and Sanz 2010; Shen et al. 2015). The IPCC AR5 reported that Spring Greenup (SG), the time at which plants begin to produce leaves in northern mid- and high-latitude ecosystems, has advanced at a rate of between 1.1 and 5.2 days per decade over different periods and regions, as inferred from multiple studies. Newer evidence indicates greater advances of SG especially over northern hemisphere high latitudes (Goetz et al. 2015; Xu et al. 2016a). Pulliainen et al. (2017) using space-borne microwave radiometer observations across northern hemisphere boreal evergreen forests for 1979–2014, report that spring recovery of carbon uptake shifted by 8.1-d (2.3 days per decade). They use this trend to estimate the corresponding changes in gross primary production (GPP) by applying *in situ* carbon flux observations. Micrometeorological $CO_2$ measurements at four sites in northern Europe and North America indicate that such advance in SG has increased the January-June GPP sum by 29 $gCm^{-2}$ (8.4 $gCm^{-2}$ (3.7%)/decade) (Pulliainen et al. 2017).

**The enhanced GPP in recent decades is the result of both longer growing season and greater greening during the growing season in the Northern latitudes, which manifest as increased seasonal amplitudes of atmospheric $CO_2$ concentrations** (Graven et al. 1960; Piao et al. 2018) (*robust evidence, high agreement*). Three satellite based leaf area index (GIMMS3g, GLASS and GLOMAP) records imply increased growing season LAI (greening) over 25–50% and browning over less than 4% of the global vegetated area, resulting in greening trend of $0.068\pm 0.045\ m^2\ m^{-2}\ yr^{-1}$ over 1982-2009 (Zhu et al. 2016). For example, GIMMS3g NDVI infers 42.0% greening and 2.5% browning of the northern vegetation from 1982 to 2014, and the greening explains 20.9% increases in the growing season productivity since 1982 (Park et al. 2016).

**The seasonal cycle of atmospheric $CO_2$ is largely driven by phenology of plant photosynthesis and ecosystem respiration. In northern ecosystems under various climate warming scenarios, additional carbon uptake attributable to enhanced photosynthetic production in summer months is greater than increased respiratory carbon emission in dormant month**s (Keenan et al. 2013; Pulliainen et al. 2017; Bond-Lamberty et al. 2018) (*robust evidence, high agreement*). The seasonal phase of atmospheric $CO_2$ level may not be easily inferred by the observed phenological shift because both photosynthesis and respiration increase during SG (Gonsamo et al. 2017). A large uncertainty exists whether the increase of soil microbial respiration with warming temperature outpaces the increase of photosynthetic productivity (Bond-Lamberty et al. 2018), and whether longer growing season may lead to moisture deficit and large year-to-year variations in annual net carbon uptake rates (Han et al. 2018; Buermann et al. 2018).

**Seasonal leaf area cycle affects both biogeochemical and biophysical interactions between vegetation and atmosphere, but varies in different regions.** Albedo exhibits seasonal patterns with development and senescence of the vegetation canopy that differentially reflect photosynthetically active and near-infrared radiation. In deciduous forests, SG increases albedo by 20–50% from the spring minima to growing season maxima, followed by rapid decrease during senescence; in contrast, in grasslands, green-up causes albedo decreases and then increases with senescence (Hollinger et al. 2010). The seasonal patterns of sensible and latent heat fluxes are also driven by LAI cycle in temperate deciduous forests: sensible heat fluxes peak in spring and autumn and latent heat fluxes peak in mid-summer (Moore et al. 1996). Increased transpiration accompanying SG causes surface cooling via latent heat transfer from land surface to the atmosphere (Schwartz, 1990). In areas with good soil water availability and extensive vegetation, increased evapotranspiration contributes enough moisture into atmosphere, resulting in increased frequency of cumulus clouds during the growing season (Richardson et al. 2013). Whereas LAI increases in warmer climate result in overall cooling effects to the climate, the greening of high latitudinal regions with vegetation may lead to positive feedback to climate, known as Arctic amplification, mainly through decrease of albedo (Pearson et al. 2013). The warming effect of albedo change is maximum in boreal summer when incoming solar radiation is high (Pielke et al. 2011; Chae et al. 2015)(*robust evidence*). In contrast, the increased growing-season greening on Tibetan Plateau has induced dominant evaporative cooling in daily maximum temperature (Shen et al. 2015). Globally, the continuously increased growing season LAI since 1982 mitigated 12% (0.09±0.02°C) of global land-surface warming for the past 30 years, via the combined cooling effects from increased evapotranspiration (70%), changed atmospheric circulation (44%) and decreased shortwave transmissivity (21%), and warming effects from increased longwave air emissivity (-29%) and decreased albedo (-6%) (Yang et al. 2017; Zeng et al. 2017).

### 2.2.5 Coupling of water in soil-plant-atmosphere continuum and drought mortality

**There is *high confidence* that hydrological response of plants to changing environmental drivers, such as soil moisture and vapour pressure deficit, mediates ecosystem responses to climate and land-atmosphere interactions** (Sellers et al. 1996; Bonan 2008). When stomatal conductance and leaf area change, the stand-level fluxes of carbon and water and latent and sensible heat fluxes also change (Seneviratne et al. 2018). Stomatal response to environmental conditions has been studied for decades (Wong et al. 1979) and has been incorporated into photosynthesis component of ESMs (Farquhar 1989). A critical role of plant water transport through the soil-plant-atmosphere continuum, particularly during periods of high temperature and drought, is widely recognised (Sperry and Love 2015; Brodribb 2009; Choat et al. 2012). New models link plant water transport with canopy gas exchange and energy fluxes, leading to improved predictions of climate change impacts on forests and land-atmosphere interactions (Wolf et al. 2016a; Sperry et al. 2017; Anderegg et al. 2016).

**Since AR5, there is limited and region specific evidence for widespread tree mortality triggered directly by drought and heat stress, often exacerbated by insect outbreak and fire** (Allen et al. 2010; Breshears et al. 2005; Kurz et al. 2008). For example, the massive climate-driven bark beetle outbreak in western Canada in the early 2000s may have converted a large region of Canadian boreal forest from a net carbon sink to a carbon source for over a decade (Kurz et al. 2008). Tree loss also alters albedo, roughness, and other biophysical properties of forests, often in complicated ways (Anderegg et al. 2012).

**There is much uncertainty in the ability of current vegetation and land surface models to adequately capture tree mortality and the response of forests to climate extremes like drought** (Hartmann et al. 2018a). Most vegetation models use climate stress envelopes or vegetation carbon balance estimations to predict climate-driven mortality and loss of forests (McDowell et al. 2011); these may not adequately project biome shifts and impacts of disturbance in future climates. For example, a suite of vegetation models was compared to a field drought experiment in the Amazon on mature rainforest trees and all models performed poorly in predicting the timing and magnitude of biomass loss due to drought (Powell et al. 2013). More recently, the loss of water transport in tree xylem due to embolism (Sperry and Love 2015), rather than carbon starvation (Rowland et al. 2015) is receiving attention as a key physiological process relevant for drought-induced tree mortality (Hartmann et al. 2018b).

A recent meta-analysis documented that a set of plant functional traits related to tree water transport explained drought-induced tree mortality rates in many sites across the world (Anderegg et al. 2016). Large uncertainties remain around how forests recover from climate stress and interactions among drought, temperature, rising $CO_2$ concentrations and other disturbances (e.g., fire, see Cross-Chapter Box 3: Fire and Climate Change).

### 2.2.6 Soil organic matter and nutrient dynamics

**Much uncertainty remains about how global carbon cycle projections may be affected by responses of soil organic matter (SOM) stocks to changes in climate, plant productivity and microbe-mediated decomposition.** Todd-Brown et al. (2013) identified high variation in soil organic carbon (SOC) stocks among CMIP5 ESMs, with model estimates of contemporary SOC stocks ranging from 510 to 3040 Pg C. Soil microbial respiration is estimated to release 40–70 Pg C 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 Pg to a gain of 146 Pg, with differences largely explained by initial SOC stocks, differing C input rates, and different decomposition rates and temperature sensitivities (Todd-Brown et al. 2014). 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 both warming and carbon addition drive considerable uncertainty in projections of future changes in SOC stocks. The processes involving C sequestration into the soil as well as microbial community responses to warming have to be adequately understood when modelling the global carbon cycle (Singh et al. 2010).Three existing data bases (SoiGrids, the Harmonized World Soil Data Base, Northern Circumpolar Soil Database) substantially differ in estimated size of global soil carbon (SOC) stock down to 1 m depth, varying between 2500 Pg to 3400 Pg (Tifafi et al. 2018). This amount is larger than the global soil carbon stock size reported as the best estimate in AR5 WGI (1500–2400 Pg), and is four to eight times larger than the carbon stock associated with the terrestrial vegetation(Bond-Lamberty et al. 2018). Although peatlands have long been recognised as one of the most carbon rich systems (Joosten 2015), new estimates since AR5 show that much larger areas in Amazon and Congo basins are peatlans (Gumbricht et al. 2017). At the moment, estimates of global peat carbon stocks suffer from lack of data from large parts of the world, especially with respect to the depth and carbon density of peat deposits (Köchy et al. 2015). Carbon stored in permafrost and deep mineral soil layers is a substantial in addition to this.

Annually, 119 Pg C is estimated to be emitted from soil to the atmosphere, of which 50% is attributed to soil microbial respiration (Auffret et al. 2016; Shao et al. 2013). It is not possible yet to make mechanistic and quantitative predictions about how multiple environmental factors influence soil microbial respiration (Davidson and Janssens 2006; Sugama et al. 2013). Meta-analyses of soil warming experiments have shown significant variability in temperature and moisture responses across biomes and climates; Crowther *et al.* (2016) found that warming effects were most sensitive to initial carbon stocks, while van Gestel et al. (2018) suggested that SOC responses to warming were not significant in an expanded version of the same dataset. Studies of SOC responses to warming over time have also shown complex responses. In a multi-decadal warming experiment, Melillo et al. (2017) found that soil respiration response to warming went through multiple phases of increasing and decreasing strength, which were related to changes in microbial communities and available substrates over time. Knorr et al. (2005) and Conant et al. (2011) suggested that transient decomposition responses to warming could be explained by depletion of labile substrates, but that long-term SOC losses could be amplified by high temperature sensitivity of slowly decomposing SOC components. Overall, long-term SOC responses to warming remain uncertain: "*although it is well established that, within reasonable limits, the biological processes which drive decomposition will be more rapid at greater temperatures, being able to assign a thermal coefficient or set of coefficients to decomposition and nutrient mineralisation has proved remarkably difficult* (Davidson and Janssens 2006; Sugama et al. 2013)". Thus, in the absence of a commonly accepted and broadly validated concept to describe SOM decomposition, projections of the impact of climate change on SOC by process-based terrestrial ecosystem models remain uncertain.

Soil moisture plays an important role in SOM decomposition by influencing microbial processes, though the exact mechanism involved is not well understood (Moyano et al. 2013). It is believed that increased soil moisture lowers C mineralisation rates under anaerobic conditions resulting in enhanced C stocks, but experimental analyses have shown that this effect may last for only 3–4 weeks after which iron reduction can actually accelerate loss of previously protected organic C by facilitating microbial access (Huang and Hall 2017). A global meta-analysis found historical constraints with moisture sensitivity of microbial respiration increasing across a precipitation gradient, resulting in about a two-fold greater loss of C from soils which were historically wetter as compared to the drier ones (Hawkes et al. 2017).

While current ESM structures mean that increasing C inputs to soils drive corresponding increases in SOC stocks, long-term carbon addition experiments have found contradictory SOC responses. Some litter addition experiments have observed increased SOC accumulation (Lajtha et al. 2014a; Liu et al. 2009), while others suggest insignificant SOC responses (Lajtha et al. 2014b; van Groenigen et al. 2014). Microbial dynamics are believed to have an important role in driving complex responses to C additions. The addition of fresh organic material can accelerate microbial growth and SOM decomposition via priming effects (Kuzyakov et al. 2014; Cheng et al. 2013). Priming effects in the soil directly surrounding living roots (rhizosphere) have been shown to increase under elevated $CO_2$ and N-limited conditions due to acceleration of root exudate production, contributing to accelerated soil C and N cycling as a plant-mediated response (Phillips et al. 2011; E. et al. 2011).

SOM cycling is dominated by “hot spots” including the rhizosphere as well as areas surrounding fresh detritus (robust evidence) (Finzi et al. 2015; Kuzyakov and Blagodatskaya 2015). This complicates projections of SOC responses to increasing plant productivity; increasing C inputs could promote higher SOC storage, but these fresh C inputs could also deplete SOC stocks by promoting faster decomposition (Hopkins et al. 2014; Guenet et al. 2018; Sulman et al. 2014). A meta-analysis by van Groenigen et al. (2014) suggested that elevated $CO_2$ accelerated SOC turnover rates across several biomes. These effects could be especially important in high-latitude regions where soils have high organic matter content and plant productivity is increasing (Hartley et al. 2012), but have also been observed in the tropics (Sayer et al. 2011).

**Microbial physiology, in particular acclimation (*via* physiological or community-level changes) to changing temperature regimes, is a large source of uncertainty in predicting SOC dynamics under warming** (Bradford et al. 2008; Zhou et al. 2012) (*medium evidence*, *high agreement*). However, experimental studies of microbial acclimation to warming have found contradictory results (Luo et al. 2001; Carey et al. 2016) with no acclimation observed in C-rich calcareous temperate forest soils (Schindlbacher et al. 2015) and arctic soils (Hartley et al. 2008). Indeed, research on soils from a variety of ecosystems from the Arctic to the Amazon indicated that microbes, in fact, could enhance the temperature sensitivity of soil respiration in Arctic and boreal soils, thereby releasing even more carbon than currently predicted (Karhu et al. 2014). In tropical forests, P limitation of microbial processes is a key factor influencing soil respiration (Camenzind et al. 2018). Temperature responses of symbiotic mycorrhizae differ widely among host plant species, without a clear pattern that may allow generalisation across plant species and vegetation types (Fahey et al. 2016).

**Temperature responses of microbial parameters, such as carbon use efficiency and soil N dynamics, have large influence on SOC responses to warming** (Allison et al. 2010; Frey et al. 2013; Wieder et al. 2013; García-Palacios et al. 2015) (*robust evidence*, *high agreement*). More complex community interactions including competitive and trophic interactions could drive unexpected responses to SOC cycling to changes in temperature, moisture, and C inputs (Corrado and Cantatore 2005; Buchkowski et al. 2017). Competition for nitrogen among bacteria and fungi could also suppress decomposition (Averill et al. 2014). Overall, the roles of soil microbial community and trophic dynamics in global SOC cycling remain very uncertain.

**Along with biological decomposition, another source of uncertainty in predicting responses of SOC to climate change is stabilisation via interactions with mineral particles (**Kleber et al., 2011; Kögel-Knabner et al., 2008; Marschner et al., 2008; Schmidt, 2011) (*high agrement*). Historically, conceptual models of SOC cycling have centred on the role of chemical recalcitrance, the hypothesis that long-lived components of SOC are formed from organic compounds that are inherently resistant to decomposition. Under the emerging new paradigm, stable SOC is primarily formed by the bonding of microbially-processed organic material to mineral particles, which limits the accessibility of organic material to microbial decomposers (Kallenbach et al. 2016; Kleber et al. 2011; Hopkins et al. 2014). SOC in soil aggregates can be protected from microbial decomposition by being trapped in soil pores too small for microbes to access (Blanco-Canqui and Lal 2004; Six et al. 2004a) or by oxygen limitation (Keiluweit et al. 2016). Alternatively, organic materials can be stabilised through chemical bonds with mineral surfaces and metal ions (Lützow et al. 2006). These organo-mineral bonds are highly stable and are thought to make bonded organic matter inaccessible to microbial decomposition, although there is some evidence that root exudates such as oxalic acid can release mineral-associated organic matter (Keiluweit et al. 2015). While some emerging models are integrating these mineral protection processes into SOC cycling projections (Wang et al. 2012; Sulman et al. 2014; Tang and Riley 2015; Wieder et al. 2015b), the sensitivity of mineral-associated organic matter to changes in temperature, moisture, fire (see Box 2.1) and carbon inputs is highly uncertain.

Deep soil layers (below 30 cm) can contain much more carbon than previously assumed (*medium evidence, medium agreement*). Based on radiocarbon measurements, deep SOC can be very old, with residence times up to several thousand years (Rumpel and Kögel-Knabner 2011) or even several tens of thousands of years (Okuno and Nakamura 2003). More recently, Strey et al., (2017) show that in deep Amazon oxisols, only 21% of the soil carbon occur in the top 0.3 m (the depth considered in the standard IPCC protocol and UNFCC guidelines) of the vertical soil profile, whereas 84% of soil carbon can be accounted by going down to 3 m. Dynamics associated with such deeply buried carbon remain understudied and ignored by the models, and not addressed in most of the studies assessed in this subsection. Deep soil C is thought to be stabilised by mineral interactions, but recent experiments suggest that $CO_2$ release from deep soils can also be increased by warming, with a 4˚C warming enhancing annual soil respiration by 34–37% (Hicks Pries et al. 2017) or additions of fresh carbon (Fontaine et al. 2007). While erosion is not typically modelled as a carbon flux in ESMs, erosion and burial of carbon-containing sediments is likely a significant carbon transfer (Zádorová et al. 2011; Asefaw et al. 2008; Wang et al. 2017d) (*medium evidence, high agreement).*

### 2.2.7 Agricultural land management and climate

**Agricultural activities impact land-climate interactions not only through land cover changes, but also through changes in land management under intensification achieved with new technology and, on the other hand, through land abandonment** (*robust evidence, high agreement*) (2.4, Chapter 4). During the 20th century, total global crop production nearly tripled, with major crop breeding improvements specifically targeting maize, rice, wheat, and other major grains (and later on, soybean) (Pingali 2012). By the late 20th century, these crops accounted for about 63% of the total global cropped area, with large expansions and land conversions across the developing world. Agriculture now occupies about 38% of the Earth's land surface (Foley et al. 2011; Ramankutty et al. 2008) (Figure 2.1). Demands for global commodity such as oil palms, rubber and soy beans, is a major driver of forest cover loss in tropical America and Asia (Curtis et al. 2018). Alongside these land conversions were increased use of chemical fertilisers and irrigation infrastructure to increase agricultural productivity, which resulted in substantial climate and environmental impacts (*high evidence, high agreement*) (Campbell et al. 2017; Pingali 2012; Foley et al. 2011)(Gleeson et al. 2012, Figure 2.3).

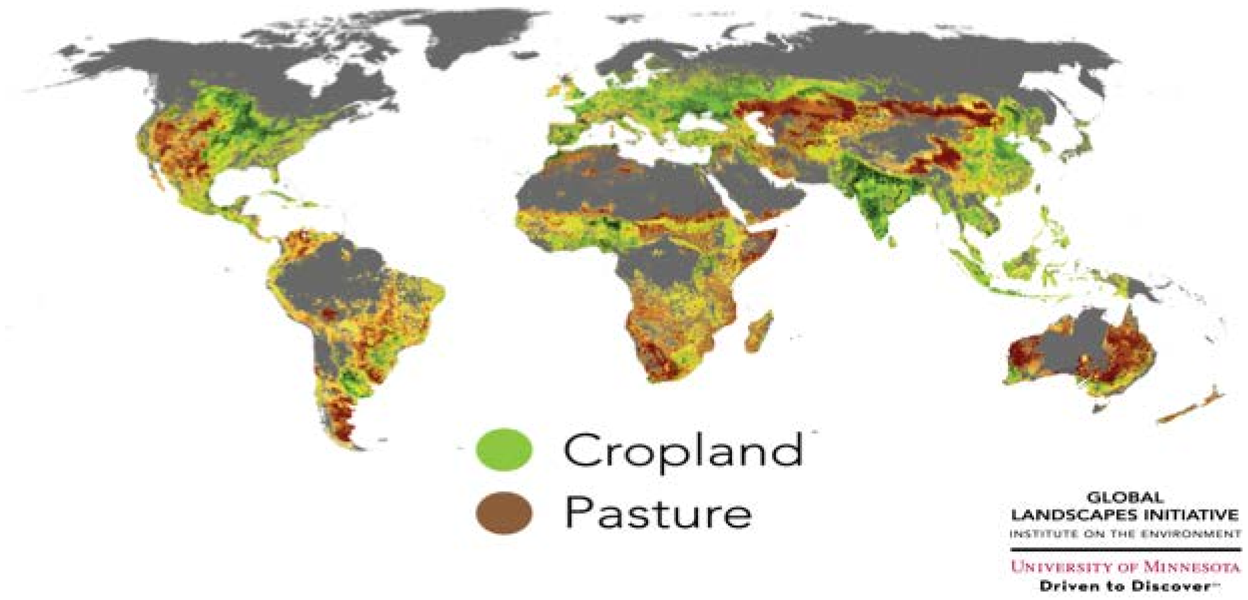


**Figure 2.3 Extent of crop and pasture areas**

Enteric fermentation associated with livestock production and rice paddy cultivation are major drivers of agricultural $CH_4$ emissions, while $N_2O$ emission largely result from agricultural soils, fertiliser applications, and manure management (Smith et al. 2008; Carlson et al. 2016). There are still outstanding uncertainties in the biogeophysical climate impacts of various agricultural land management strategies and the response of managed vegetation to global environmental change. Improvements in observational datasets, however, are now elucidating important trends and processes by which agricultural land management is impacting regional and global climate systems (Duveiller et al. 2018a). For example, a non-trivial amount (about 20%) of land carbon sink strength may also be explained by intensifying agricultural production trends that involve the prodigious use of nitrogenous fertilisers and irrigation (Mueller et al. 2014). These trends are most pronounced in areas of rapid agricultural LCC and development over this time period, such as those in South and East Asia, as well as portions of Europe and South America (Liu et al. 2015a; Zeng et al. 2017). More research is needed to disentangle the effects of intensive management also embedded in observed greening signals.

**There is *medium evidence and medium agreement* that $CO_2$ fertilisation effects have increased water use efficiency and thus reduced agricultural water per unit amount of crop produced** (Deryng et al. 2016; Nazemi and Wheater 2015; Elliott et al. 2014). This effect may be quite pronounced in semi-arid and arid environments, which could lead to near-term continued greening of agricultural areas alongside modifications to water management, however current assessments of these effects are based on limited datasets from mostly temperate growing regions (Deryng et al. 2016).

There is *medium confidence* in the extent to which agriculture-driven soil fertility loss and soil degradation affect local and regional hydro-climate (Amundson et al. 2015; Lal 2011). Agricultural soils in some of the most productive regions have exhibited carbon losses due to include ploughing and tillage, over-fertilisation, and disappearance of long fallow-periods (Arneth et al. 2017; Pugh et al. 2015b; Lal 2011). These practices may have resulted in soil organic carbon losses ranging from 25–75% across global agricultural regions (Sanderman et al. 2017; Lal 2011), although much uncertainty still exists (Pongratz et al. 2014) (*medium confidence, medium evidence*). The removal of organic matters can also impact the soil's capacity to store and filter water throughout the column and within the root zone (Amundson et al. 2015), but the magnitudes of these effects on climate processes remain uncertain (Minasny and McBratney 2018).

Emerging land management options for mitigation of climate impacts include deliberately planned crop rotations, timing, and water/irrigation(Hirsch et al. 2017; Seneviratne et al. 2018). Additionally, regionally degraded agricultural soils could potentially serve as carbon sinks by implementation of improved nutrient and water management techniques that enable high agricultural productivity and organic matter return(Bustamante et al. 2014; Paustian et al. 2016; Minasny et al. 2017). However, outstanding uncertainties exist in quantifying the amount of carbon that can be stored locally and regionally, which depend on the confluence of biogeophysical, biogeochemical, and socio-economic conditions (Powlson et al. 2014b).

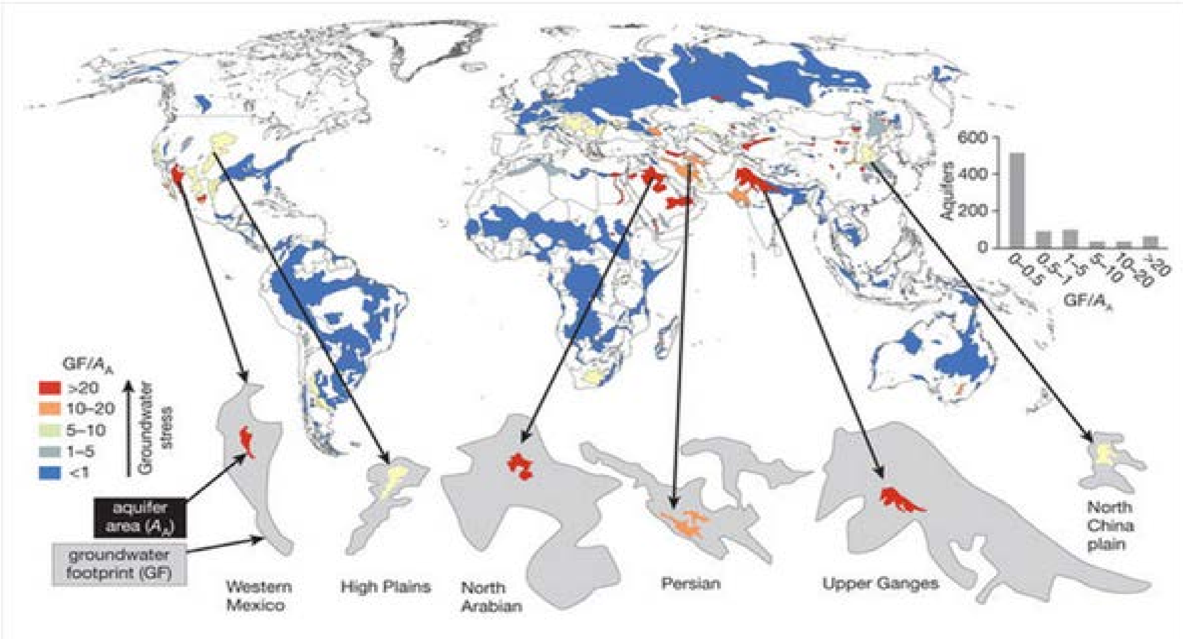


**Figure 2.4 Groundwater footprint of major aquifers used for agricultural irrigation. Higher values indicate stressed conditions. Adapted from (Gleeson et al. 2012)**

***Climate patterns and processes related to agricultural irrigation***

**An increasing body of climate modeling work demonstrates that intensive irrigation potentially exerts a strong climate forcing** (Cook et al. 2015b; Guimberteau et al. 2012) (*high confidence, robust evidence*). Nearly 70% of global freshwater withdrawals, approximately 3300 km$^3$yr$^{-1}$ in 2010, are currently used for agricultural irrigation with groundwater accounting for about 30–40% of this total (Pokhrel et al. 2016; Wada et al. 2014). In some regions, such as South Asia, year-round irrigation consumes about 90% of freshwater withdrawals from surface and groundwater stores combined (Dominik et al. 2008; Gleeson et al. 2012; Thakur and Jayangondaperumal 2015). The most intensive irrigation regimes may be found in water-limited regions, where about 45% of global agricultural productivity takes place (Pokhrel et al. 2016). Addition of such vast amounts of water to the land surface can substantially modify regional energy and moisture balances, particularly in conjunction with highly productive agricultural crops with high rates of evapotranspiration. In general, climate studies and assessments of irrigation have sought to quantify and understand how irrigation-induced enhancements in surface latent heat fluxes can impact overall regional energy and moisture balances and interact with larger-scale atmospheric circulation processes, particularly in water-limited domains. However, most CMIP5 models did not account for water management.

### 2.2.8 Urbanisation and climate change

**Urbanisation impacts local climates surrounding urban areas** (Wang et al. 2016b; Zhong et al. 2017). Globally, urbanisation *per se* is not a direct driver of forest loss (Curtis et al. 2018), but energy and resource demands in urban areas drive global trades and indirectly influence land-climate interactions (2.2.7). **Climate change in urban areas affect health and energy demand of large number of people living in urban areas** (Li et al. 2017b; Santamouris et al. 2015) (*robust evidence, high agreement*). Urban heat island (UHI), with temperature in urban areas higher than that in the surrounding rural areas, has intensified as anthropogenic heat discharges have increased, whereas albedo and vegetation coverage have decreased

(Mohajerani et al. 2017; Phelan et al. 2015). With the time series of satellite observations of land surface temperature, surface UHI (SUHI), especially its trend (REBATTU and DUPOUX 1945; Zhou et al. 2016) and regional variations (Zhou et al. 2017) have been investigated. The intensity of SUHI varies across regions and study areas, for example, less than 0.5°C in Mediterranean cities (Polydoros et al. 2018) and higher than 8°C in Baguio City, Philippines (REBATTU and DUPOUX 1945).

Urbanisation alters the stock size of soil organic carbon (SOC) and its stability by converting natural vegetation to urban land cover. Overall, carbon densities or stocks decreases from natural land areas to urban core along the rural-urban gradient (Tao et al., 2015; Zhang et al., 2015). The conversion of vegetation, to urban land results in a loss of carbon stored in plants, and stresses associated with urban environment (e.g., heat, limited water availability and pollution) may reduce plant growth and survival (Xu et al. 2016b). However, urban soils may serve as an important carbon sink in some areas (Yesilonis et al. 2017). Urban soils may exhibit high levels of SOC, significant enough to be considered in earth system models (Zhai et al. 2017; Vasenev et al. 2018). Urbanisation *per se* may not result in loss of carbon already present in the soil (Liu et al. 2018a), although there is a wide variation across different urban landscapes. For example, a tenfold difference in SOC stock across land cover types was reported for Seoul Forest Park (urban park)(Bae and Ryu 2015). In Changchun in Northeast China, SOC density is higher in recreational forests within urban areas compared to a production forest (Zhang et al. 2015).

**Urbanisation changes precipitation patterns, frequency, and intensity as a results of changes in thermodynamic, aerodynamic, and cloud microphysics** (Zhong et al. 2015, 2017) (*medium evidence, medium agreement*). Divergent results have been reported from different areas using different detection methods. Some report that high temperature of UHI increases the formation of convective clouds over urban areas, leading to an increase in the frequency of extreme summer precipitation (Shimadera et al. 2015; Dou et al. 2015; Zhong et al. 2017). A similar finding is reported based on the satellite observation of the urban area of the China Pearl River Delta; increases in short-duration heavy rain is observed but less so compared to surrounding rural areas (Chen et al. 2015). The urbanisation-induced convection may result in stronger effects on precipitation near large water bodies from which humid atmosphere may be drawn into (Kusaka et al. 2013; Yang et al. 2013). On the other hand, during the weak UHI periods, summer thunderstorms may bifurcate and bypass the urban center because of the building-barrier effect, producing a minimum of regional-normalised rainfall in the urban center and directly downwind of the urban area (Dou et al. 2015). Additionally, increased aerosols in urban areas may counter the effect of UHI on precipitation (Zhong and Yang 2015; Zhong et al. 2017) (*limited evidence*).

## 2.3 The effect of climate variability and change on land

### 2.3.1 Overview of climate impacts on land

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 Peixóto 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 Koppen-Geiger classification, Kottek et al. 2006). Vegetation biomes are adapted to these regional climates and their variability and may shift as climate, land surface characteristics (e.g., geomorphology, hydrology) and ecosystems interact (Figure 2.5). Functioning within these biomes is subject to modes of natural variability in the ocean-atmosphere system that result in wetter/dryer or hotter/cooler periods regionally having temporal scales from weeks to months (e.g., Southern Annular Mode), months to seasons (e.g., Madden-Julien 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, shape ecosystems at various space and time scales. As climate is a primary determinant of both global and regional land characteristics and functioning, climate change due to natural or anthropogenic causes can alter these. Climate change alters the drivers of natural climate variability and the characteristics of climate extremes with subsequent impacts on terrestrial ecosystems and 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).

As a result of a warming climate, tropical and sub-tropical regions will see the emergence of novel climates that are beyond the envelope of current natural variability (*high agreement, robust evidence*) (Mora et al. 2013, 2014; Hawkins et al. 2014; Colwell et al. 2008; Maule et al. 2017) and hot, arid climates are projected to expand (Section 2.3.2). It is *very likely* that many land-based systems will be exposed to disturbances beyond the range of current natural variability, which will alter the structure, composition and functioning of the system (Settele et al. 2015; Gauthier et al. 2015). It is *very likely* that changes to natural climate variability as a result of global warming, particularly through changes in extreme weather events, will impact terrestrial ecosystems including croplands and thus food security (Sections 2.3.3 and 2.3.4).

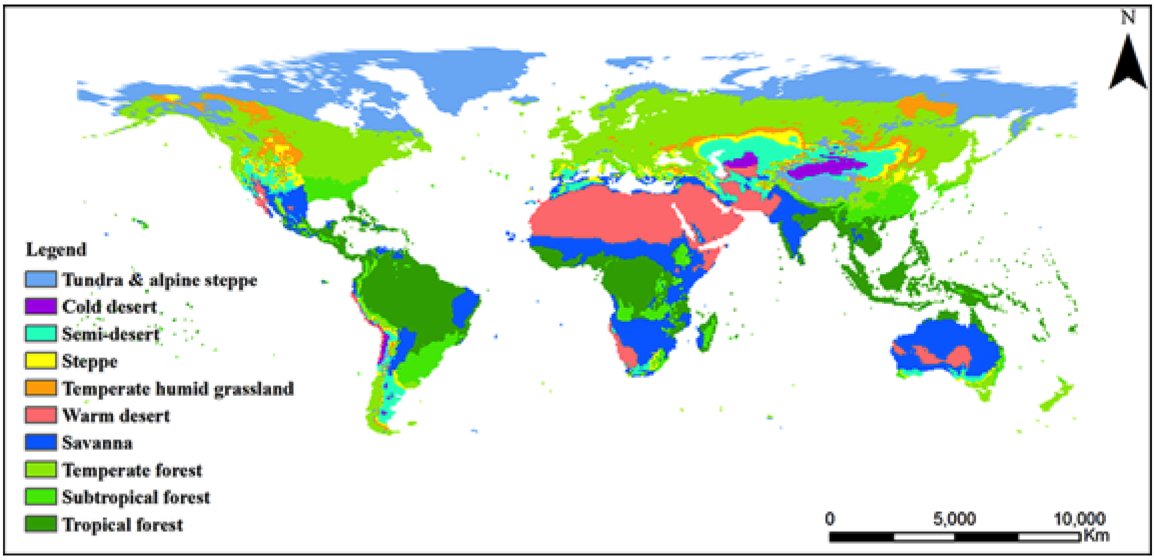


**Figure 2.5 Global natural vegetation biomes and their spatial variability during the period 1971–2000. The different colours represent the different ecosystems, tundra & alpine steppe (12.85%), cold desert (1.56%), warm desert (13.93%), semi-desert (5.96%), savanna (17.66%), steppe (3.29%), temperate humid grassland (5.86%) and Forest (39.65%) in this period. From (Gang et al. 2013)**

### 2.3.2 Desertification and land degradation

Desertification, defined and discussed at length in Chapter 3 of this report, occurs in drylands and is a function of both human activity and climate variability and change. Most drylands exist along climate ecotones (transition zones between a wet and a dry climatic regime, often with sharp gradients in precipitation), are the consequence of complex feedback loops between climatological, biological, geomorphological, hydrological, and human systems (Nicholson 2011). They are usually characterised by strong seasonal and interannual climate variability and due to the relative scarcity of precipitation in drylands, small changes in rainfall can have large impacts on these systems.

Different methods used for determining rates of desertification (Thorstensson 2001; Safriel 2007) all show that the extent of global drylands has increased over the last 60 years (Dai 2013; Feng and Fu 2013; Huang et al. 2016, 2017), although there are uncertainties in distinguishing between climate-caused and anthropogenic desertification (D'Odorico et al. 2013). Future projections show an increase in aridity across the globe, attributable to GHGs (Burke et al. 2006; Dai 2011; Rajaud and Noblet-Ducoudré 2017) and the extent of global drylands is projected to accelerate in the 21st century, with most of this expansion expected to occur in developing countries (Lickley and Solomon 2018; Huang et al. 2017). Within this net dryland expansion, temperate drylands are projected to convert to subtropical drylands as a result of an increased frequency of ecological drought in temperate drylands leading to reduced soil moisture availability in the growing season (Schlaepfer et al. 2017). Biological soil crusts that cover dryland soils and stabilise soils, influence the water cycle and vegetation growth, and contribute to the biogeochemical cycles of carbon and nitrogen are expected to shrink by 27–39% with negative impacts on soil stability and overall dryland fertility (Elbert et al. 2012; Belnap et al. 2016; Rodriguez-Caballero et al. 2018). Worryingly, dryland expansion has been underestimated in the historical simulations of the CMIP5 GCMs (Feng and Fu 2013) and Huang et al. (2016) estimate 56% and 50% of total land surface will be covered by drylands by 2100 under RCP8.5 and RCP4.5, respectively. Projected warming trends over drylands are twice the global average and a larger area of drylands are projected to dry earlier and more severely than humid areas (Lickley and Solomon 2018)**.** Along with extensive land use, the risk of land degradation and desertification affects approximately 70% of the Earth's agricultural drylands (Huang et al. 2017 and citations therin). Dryland expansion will lead to reduced carbon sequestration and enhanced regional warming, result in increased water scarcity, vegetation die-offs, decreased agricultural yields and increased drought frequency and persistence (Allen et al. 2010; Prudhomme et al. 2014; Schewe et al. 2014; Huang et al. 2017).

### 2.3.3 Climate-driven changes in terrestrial ecosystems

Previous IPCC AR5 reported *high confidence* that the earth’s biota composition and ecosystem processes have been affected strongly by past changes in global climate, at a climate change rate lower than those projected for the 21st century under high warming scenarios like RCP8.5 (Settele et al. 2015). Moreover, most ecosystems are vulnerable to climate change at projected rates of warming even under low- to medium-range warming scenarios. There is *high confidence* that as a result of climate changes over recent decades many plant and animal species have experienced changes in sizes and locations of ranges, altered abundances, and shifts in seasonal activities (Feeley et al. 2011; Ernakovich et al. 2014; Elsen and Tingley 2015; Hatfield and Prueger 2015; Urban 2015; Savage and Vellend 2015; Yin et al. 2016; Pecl et al. 2017).

In a warming climate many species will be unable to track their climate niche as it moves, especially those in extensive flat landscapes and with low dispersal capacity (Warszawski et al. 2013) and the tropics whose thermal optimum is already near current temperature. While climate change will be the principal driver of range contractions at higher latitudes, land conversion (e.g., deforestation, conversion of grasslands to croplands, etc.) will have a much larger effect on species that inhabit the tropics (Jetz et al. 2007). Expansion of forest at higher elevations occurs as a result of abandoned land use and climate change and may favour thermophilic species at the expense of cold adapted species (Grace et al. 2002; Harsch et al. 2009; Landhäusser et al. 2010; Alatalo and Ferrarini 2017; Rumpf et al. 2018; Steinbauer et al. 2018). Nonetheless, these effects can be countered by intense and frequent drought conditions which result in accelerated rates of taxonomic change and spatial heterogeneity in an ecotone (Tietjen et al. 2017).

Since the advent of satellite observation platforms, a global increase in vegetation photosynthetic activity (greening) as evidenced through remotely sensed indices such as leaf area index (LAI) and normalised difference vegetation index (NDVI) has been observed (Myneni et al. 1997; de Jong et al. 2012; Los 2013; Piao et al. 2015; Mao et al. 2016; Zhu et al. 2016; Carlson et al. 2017). This greening has occurred in the Amazonia, parts of Australia, the Sahel, India, China (Piao et al. 2015) and the northern extratropical latitudes (Pan et al. 2018) and is attributable primarily to $CO_2$ fertilisation, which explains most greening trends in the tropics, and to a lesser degree by climate change through an extended growing season, particularly in the high latitudes and Tibetan Plateau, and nitrogen deposition (Fensholt et al. 2012; Zhu et al. 2016). Within the global greening trend are also detected regional decreases in vegetation photosynthetic activity (browning) in northern Eurasia, the southwestern USA, boreal forests in North America, Inner Asia and the Congo Basin, largely as a result of intensified drought stress. Since the late-1990s rates and extents of browning have exceeded those of greening, and areas experiencing greening-to-browning reversals are much larger than those experiencing browning to greening trends (de Jong et al. 2012), which has resulted in a slowdown of the global greening rate. Projections of greening/browning trends are uncertain. Projected increases in drought conditions in many regions suggest global vegetation greening trends are at risk of reversal to browning in a warmer climate (de Jong et al. 2012; Pan et al. 2018) but in higher latitudes vegetation productivity is projected to increase as a result of higher atmospheric $CO_2$ concentrations and longer growing periods (Ito et al. 2016)(Section 2.4).

Increased $CO_2$ concentrations in the atmosphere has both direct and indirect effects on terrestrial ecosystems (see 2.2.2 and 2.2.3). The direct effect is primarily through increased vegetation photosynthetic activity as described above. Indirect effects include decreased evapotranspiration that may offset the projected impact of drought in some water-stressed plants through improved water use efficiency in temperate regions suggesting that rain-fed cropping systems will benefit from elevated atmospheric $CO_2$ concentrations (Roy et al. 2016; Milly and Dunne 2016; Swann et al. 2016; Zhu et al. 2017a). In tropical regions increased flowering activity is associated primarily with increasing atmospheric $CO_2$ suggesting a long-term increase in flowering activity may persist in some vegetation, particularly midstory trees and tropical shrubs, and enhance reproduction levels until limited by nutrient availability or climate factors like drought frequency, rising temperatures and reduced insolation (Pau et al. 2018).

### 2.3.4 The influence of climate change on food security

Food security (see Chapter 5 and glossary) is a function of climatic factors (temperature, rainfall, $CO_2$, ozone), non-climate factors (soil fertility, irrigation, demography, economics), production factors (crops and livestock) and non-production factors (processing, transport, storage, retail, income). Therefore the overall impact of climate on food security is complex, being greater than impacts on agricultural productivity alone (Vermeulen et al. 2012a; Groppo and Kraehnert 2017).

Climate change is expected to substantially affect the sustainability of food availability, access, and utilisation, leading to complex impacts on global food security. With respect to availability, specifically agricultural production, climate change will have regionally-distributed impacts, even under aggressive mitigation scenarios (Millar et al. 2014; Rosenzweig et al. 2013; Challinor et al. 2014; Parry et al. 2005; Lobell and Tebaldi 2014; Wheeler and Von Braun 2013). At middle and higher latitudes, the lengthening of growing seasons, reduced frost damage, $CO_2$ fertilisation effects, potential for increased rainfall and northward expansion of warmer temperatures and conducive climate conditions (Gregory and Marshall 2012; Yang et al. 2015b) may benefit crop productivity (Parry et al. 2004; Rosenzweig et al., 2014; Deryng et al. 2016) (*medium confidence, medium evidence*). However, recent climate trends' negative impacts on grain yields (through enhanced climate variability and extremes and nutrient limitation) together with non-climate factors such as soil health and competitive plant species generally outweigh the positive impacts globally, including some middle to higher latitudes countries (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).

The negative impacts of climate change on crop yields may be particularly pronounced in the sub-tropics, tropics, and water-limited environments as rainfall variability increases, drought severity is enhanced, and growing season temperatures rise (Parry et al. 2004; IFPRI 2009; Schlenker and Lobell 2010; Müller 2011; Challinor et al. 2014; Müller et al. 2017). In addition to these direct climate impacts on agriculture, indirect climate impacts such as reductions in soil fertility and increased erosion (Jobbins and Henley 2015), the expansion of geographic ranges and/or relocations of crop pests and diseases and modifications of their development and lifecycles (Tirado et al. 2010; Jobbins and Henley 2015a; Moses et al. 2015; Fanzo et al. 2018) and competition with agricultural weeds may also decrease crop productivity and/or enhance crop losses, particularly in warmer climates. Such impacts could generate geographic shifts in land use and crop distributions for some regions (Ziska, L. H., and Runion 2006; Rosenzweig and Tubiello 2007).

Over 60% of the world's crop production is dominated by maize, rice, wheat and soybean (Pingali 2012). Analyses of historical crop production trends indicate climate-induced 20th century reductions in overall global wheat and maize yields, while rice and soybean responses showed much regional variation (Zhao et al. 2017; Lobell et al. 2011). In the absence of adaptation, many crops may exhibit global yield declines and enhanced variability after increases of 2˚C in globally averaged temperature (or by mid-century under high GHG emissions trajectories) (Vermeulen et al. 2012a; Challinor et al. 2014; Deryng et al. 2014; Lobell and Tebaldi 2014) (*high confidence, medium evidence*). However, there exists outstanding uncertainty due to the magnitude of $CO_2$ fertilisation effects, which could partially mitigate mean yield losses in higher latitude regions (*medium confidence, medium evidence*) (Deryng et al. 2016). 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) (*medium confidence, limited evidence).* Temperature increases are also expected to be a constraining factor for maize productivity by the end of the century, resulting in potential yield reductions across both mid- and low latitude regions (Bassu et al. 2014; Zhao et al. 2017) and climate shocks, particularly severe drought, are expected to impact low-income small-holder producers disproportionately (Lopez-i-Gelats 2014).

Several sources of uncertainty exist in these projected climate change-induced crop impacts, partly stemming from differences between the utilised tools and models, sparse observations to current climate trends, and other agro-ecosystem responses (e.g., to $CO_2$ effects) (Asseng et al. 2013; Li et al. 2015b; Mistry et al. 2017; Bassu et al. 2014). The uncertainty in climate simulation is generally larger than, or sometimes comparable to, the uncertainty in crop simulations using a single model (Iizumi et al. 2011), but is less than crop model uncertainty when multiple crop models are used and $CO_2$ is considered (Koehler et al. 2013; Müller et al. 2015; Hasegawa et al. 2017; Asseng et al. 2013).

Climate-change induced temperatures increases, water resource variability, extreme events, and the propagation of diseases, are also likely to negatively impact the livestock sector, particularly with respect to animal growth, health, reproduction, and milk production (Rojas-Downing et al. 2017; Thornton et al. 2009). While opportunities may exist for improved forages due to $C_3$ plant responses under higher $CO_2$ concentrations, overall climate change impacts on the productivity, digestibility, and nutrient quality of feed crops could put further pressures on livestock-oriented agro-ecosystems (Polley et al. 2013; Rojas-Downing et al. 2017; Thornton et al. 2009) (*medium agreement, medium evidence*). Increases in the geographic extent and prevalence of livestock diseases and the prevalence of disease vectors under warmer conditions and longer growing seasons may pose substantial risk to livestock systems (Nardone et al. 2010; Rojas-Downing et al. 2017) (*medium agreement, limited evidence*). Higher temperatures and humidity levels may also incur greater risks of livestock heat stress that induce increased water demand, reduce weight gain and milk production (or eggs in the case of poultry), although these thresholds vary by animal species (Nardone et al. 2010).

The impacts of climate change also extend far beyond production, affecting post-harvest food storage and critical supply chain components, thereby challenging food accessibility and utilisation, while also contributing to losses and wastes (Fanzo et al. 2018). For example, higher temperature and humidity levels may present risks to post-harvest, stored grain and foodstuffs through spoilage, and by providing environments conducive to the development of microorganisms, mycotoxins, and other contaminants (Vermeulen et al. 2012b; Battilani et al. 2016; Fanzo et al. 2018; Moses et al. 2015). Sea level rise, as well as changes in the frequency and severity of extreme storms, and other climate events, may 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; Vermeulen et al. 2012b). When taken together with increased post-harvest spoilage, pests, and contamination risks, such climate impacts could spur additional upstream losses and downstream wastes throughout the food supply chain and system.

These direct and indirect climate change impacts could drive increased adoption of cold chains, which allow produced, perishable foodstuffs to be refrigerated for extended periods of time before market transactions thereby increasing costs and the emissions of GHG (Coulomb 2008; Vermeulen et al. 2012; Porter and Reay 2016; Fanzo et al. 2018). Importantly, food access within countries relies not just on local or domestic production, but also on imports and distribution networks, which may be impacted by both climate change and climate variability (Lopez-i-Gelats 2014; Bailey and Wellesley 2017). Climate changes and subsequent effects on trade networks and physical infrastructure may limit the import of critical foodstuffs and/or impact pricing and scarcity, leading to regional food insecurity and instability. These impacts may be particularly acute for staple grains, for example rice and wheat, which have relatively complex trade connections between a few exporting and many importing states (Müller 2011; Nelson et al. 2014; Puma et al. 2015). Local and regional declines in crop productivity due to climate change and extremes, in combination with potential climate induced failures in trade and distribution networks, may also contribute to human migration and regional conflict, thereby further dislocating vulnerable populations from food sources (Selby et al. 2017; Kelley et al. 2015; Challinor et al. 2018).

### 2.3.5 Climate extremes and their impact on land type and functioning

Extreme climate events are generally defined as the upper or lower ends of the observed range of values of climate variables or climate indicators (e.g., temperature/rainfall or drought/aridity indices respectively). Extreme events occur across a wide range of time and space scales and may include individual, relatively short-lived weather events (e.g., extreme thunderstorms storms, frost events) or result from an accumulation of non-extreme climate events (e.g., floods, heat waves and drought). Recent IPCC reports have reported with *high confidence* on the increase of many types of observed extreme temperature events (Seneviratne et al. 2012; Hartmann et al. 2013). However, as a result of observational constraints, increases in precipitation extremes are less confident, except in observation rich regions with dense, long-lived networks such as Europe and North America where there have been *likely* increases in the frequency or intensity of heavy rainfall.

Extreme climate events typically have extreme impacts on land type and functioning at different scales in space and time (Figure 2.6). Extreme impacts may also arise through a combination of variables that are not necessarily in an extreme state (Colwell et al. 2008; Kundzewicz and Germany 2012). These combinatory processes leading to a significant impact are referred to as a compound event and are a function of the nature and number of physical climate and land variables, the range of spatial and temporal scales, the strength of dependence between processes, and the perspective of the stakeholder who defines the impact (Leonard et al. 2014). Current *confidence* in the impact of compound events on land functioning and type is *low* as the multi-disciplinary approaches needed to address the problem are few (Zscheischler et al. 2018).

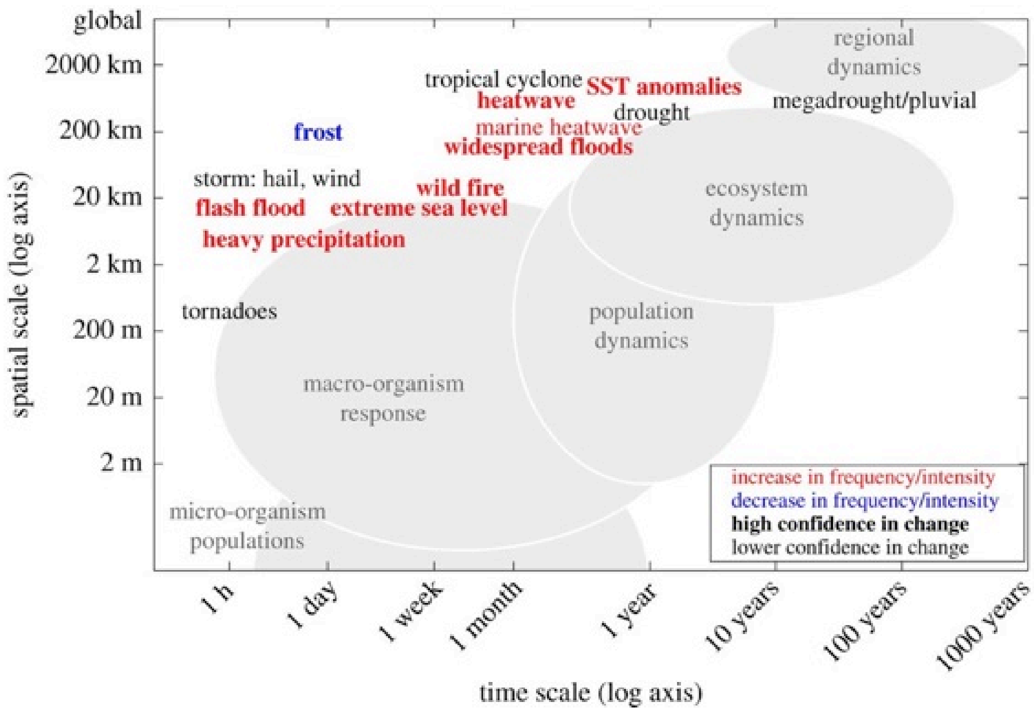


**Figure 2.6 Spatial and temporal scales of typical extreme climatic events (ECEs) and the biological systems they impact (shaded grey). Individuals, populations and ecosystems within these space-time ranges respond to relevant climate stressors. Red (blue) labels indicate an increase (decrease) in the frequency or intensity of the event, with bold font reflecting confidence in the change. For each ECE type indicated in the figure, ECEs are likely to affect biological systems at all temporal and spatial scales located to the left and below the specific ECE position in the figure. From Ummenhofer and Meehl (2017)**

#### *2.3.5.1 Changes in extreme temperatures, heat waves and drought*

It is *very likely* that most land areas have experienced a decrease in the number of cold days and nights, and an increase in the number of warm days and unusually hot nights (Orlowsky and Seneviratne 2012; Seneviratne et al. 2012; Mishra et al. 2015; Ye et al. 2018). Despite there being no consensus definition of heat waves (e.g., some heat wave indices have relative thresholds and others absolute thresholds), there is significant correlation in observed heat wave trends between indices of the same type and that recent heat-related events have been made more frequent or more intense due to anthropogenic greenhouse gas emissions (Lewis and Karoly 2013; Smith et al. 2013b; Scherer and Diffenbaugh 2014; Fischer and Knutti 2015; Ceccherini et al. 2016; King et al. 2016; Bador et al. 2016; Stott et al. 2016; King 2017). Globally, 50-80% of the global land fraction is projected to experience significantly more intense hot extremes than historically recorded (Fischer et al. 2013; Diffenbaugh et al. 2017; Seneviratne et al. 2016). There is *high confidence* that heat waves will increase in frequency, intensity and duration into the 21st century (Russo et al. 2016; Ceccherini et al. 2017; Herrera-Estrada and Sheffield 2017) and by the end of the century heat waves may become extremely long (more than 60 consecutive days) and frequent (once every two years) in Europe, North America, South America, Africa, Indonesia, the Middle East, south and south east Asia and Australia (Rusticucci 2012; Cowan et al. 2014; Russo et al. 2014; Scherer and Diffenbaugh 2014; Pal and Eltahir 2016; Rusticucci et al. 2016; Schär 2016; Teng et al. 2016; Dosio 2017; Mora et al. 2017; Dosio et al. 2018; Lehner et al. 2018; Lhotka et al. 2018; Lopez et al. 2018; Tabari and Willems 2018) and unusual heat wave conditions today will occur regularly by 2040 under the RCP 8.5 scenario (Russo et al. 2016).

The intensity of heat events may be modulated by the land cover and soil characteristics (Miralles et al. 2014; Lemordant et al. 2016). Temperature increases result in decreased soil moisture, which reduces latent heat flux, allowing temperatures to rise further, however, this feedback may be diminished if the land surface is irrigated through the introduction of latent heat (Mueller et al. 2015; Siebert et al. 2017)(Section 2.6.2.2.1). Drought is defined by the IPCC as “A period of abnormally dry weather long enough to cause a serious hydrological imbalance" recognizing that "Drought is a relative term, therefore any discussion in terms of precipitation deficit must refer to the particular precipitation-related activity that is under discussion" (Planton 2013). Drought is a normal component of climate variability (Hoerling et al. 2010; Dai 2011) and may be seasonal, multi-year (Van Dijk et al. 2013) or multi-decadal (Hulme 2001) with increasing degrees of impact on the regional activities. Droughts impact many aspects of land functioning and type including agricultural productivity (Lesk et al. 2016), 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. 2015; Doughty et al. 2015; Schlesinger et al. 2016). Although systems may demonstrate resilience to a climate stressor like drought, the compound effect of deforestation, fire and drought potentially lead to losses of carbon storage, changes in regional precipitation patterns and river discharge and a transition to a disturbance-dominated regime (Davidson et al. 2012). Additionally, adaptation to seasonal drought may be overwhelmed by multi-year drought (Brando et al. 2008; da Costa et al. 2010).

Historical droughts have largely been a function of multi-decadal to annual natural variability in the climate system, particularity through remote SST forcings such as the Inter-decadal Pacific Oscillation (IPO) and the Atlantic Multi-decadal Oscillation (AMO), El Niño/Southern Oscillation (ENSO) and Indian Ocean Dipole (IOD), that cause drought as a result of reduced rainfall (Kelley et al. 2015; Dai 2011; Hoell et al. 2017). In some cases however, large scale SST modes do not fully explain the severity of drought, for example the drought event in the western Amazon during the 2015/2016 El Niño, suggesting factors such as anthropogenic warming and land cover change need to be considered (Erfanian et al. 2017). It is currently uncertain how these large-scale modes of variability will respond to a warming climate (Deser et al. 2012; Liu 2012; Christensen et al. 2013; Hegerl et al. 2015; Newman et al. 2016), although there is evidence for an increased frequency of extreme ENSO events, such as the 1997/1998 El Niño and 1988/89 La Niña (Cai et al. 2014, 2015) and extreme positive phases of the IOD (Christensen et al. 2013; Cai et al. 2014) and potentially shorten the Pacific Decadal Oscillation (PDO) period (Zhang and Delworth 2016).

Long-term global trends in drought are difficult to determine because of this as well as potential deficiencies in drought indices (especially in how evapotranspiration is treated), discrepancies in and the availability of precipitation data and the role of natural variability (Sheffield et al. 2012; Dai 2013; Trenberth et al. 2014; Nicholls and Seneviratne 2015; Mukherjee et al. 2018). However, regional trends in frequency and intensity of drought are evident in several parts of the world, particularly in low latitude land areas, such as the Mediterranean and Middle East (Vicente-Serrano et al. 2014; Spinoni et al. 2015; Dai and Zhao 2017; Páscoa et al. 2017), West, central and parts of East and southern Africa (Dai 2013; Druyan 2011; Masih et al. 2014; Dai and Zhao 2017), Central China (Wang et al. 2017e), east and south Asia, parts of North America and eastern Australia (Dai and Zhao 2017). Although some recent detection-attribution studies have identified a climate change fingerprint in several regional droughts, for example southern Europe and the Mediterranean (Kelley et al. 2015; Wilcox et al. 2018), parts of North America (Williams et al. 2015; Mote et al. 2016), Russia (Otto et al. 2012) and Australia (Lewis and Karoly 2013), most droughts have been a consequence of current natural climate variability.

It is expected that the extra heat from global warming will exacerbate heat stress and increase potential evapotranspiration, thereby amplifying deficits in soil moisture and runoff despite uncertain precipitation changes (Ficklin and Novick 2017; Berg and Sheffield 2018; Cook et al. 2018; Dai et al. 2018) This will increase the rate of drying causing natural drought to set in quicker, become more intense, last longer, become more widespread resulting in an increased global aridity (Dai 2011; Prudhomme et al. 2014) and emergent and substantive risk (see Chapter 7). Projected drought risk over the Mediterranean, central Europe, the Amazon and southern Africa increases significantly for both 1.5°C and 2°C warming levels compared to present day and the additional 0.5°C from a 1.5°C to 2°C climate leads to significantly higher drought risk (Lehner et al. 2017) and over the USA. Southwest and Central Plains a two-degree there are elevated drought risks compared to present day under both RCP4.5 and RCP8.5 scenarios, including the risk of megadrought (Ault et al. 2014; Cook et al. 2015a; Diffenbaugh et al. 2015; Williams et al. 2015; Ault et al. 2016; Udall and Overpeck 2017).

#### *2.3.5.2 Impacts of heat extremes and drought on land*

The exposure of cropping systems to extreme heat, particularly during key growth phases such as the reproductive period, can severely reduce crop production (Gourdji et al. 2013; Jagadish et al. 2015). These adverse heat conditions have been observed to reduce crop yield in many regions of the world and will continue to do so in the future in the absence of adaptation, particularly in regions dependent on rain-fed agriculture (Durigon and de Jong van Lier 2013; Siebert et al. 2014; Trnka et al. 2014; Kimball et al. 2015; Schauberger et al. 2017; Zhang et al. 2017b; Asseng et al. 2015). Unusually hot nights are damaging to most crops (Peng et al. 2004; Wassmann et al. 2009) as are extremely high daytime temperatures which are occasionally lethal to crops (Porter and Gawith 1999; Meerburg et al. 2009). Heat stress over wheat cropping regions increased significantly in the period 1980–2010 and is as important a predictor of crop yield as drought (Semenov and Shewry 2011; Teixeira et al. 2013; Zampieri et al. 2017). Heat waves in combination with drought are common and intrinsically linked through a positive feedback (Stéfanon et al. 2014). Drought, however, has a larger detrimental effect on wheat yield than heat stress in Mediterranean countries (Zampieri et al. 2017) and affects both harvested area and yield (Lesk et al. 2016). For irrigated crops heat stress is reduced due to surface cooling (Puma and Cook 2010; Troy et al. 2015; Carter et al. 2016) and may be overestimated in irrigated regions (Siebert et al. 2017).

Heat extremes impact livestock productivity and mortality directly through causing heat stress and related impacts on health, growth and reproduction (Morignat et al. 2014), compromising adaptive response mechanisms of animals, altering the spread and prevalence of diseases (Bett et al. 2017) and indirectly through reduced availability and quality of grazing forages and feed crops (Rust and Rust 2013; Giridhar and Samireddypalle 2015). Projected increases of extreme heat events result in higher risks to livestock productivity, particularly to low-income groups dependent on subsistence agriculture and in arid and semi-arid regions (Rust and Rust 2013).

Extreme heat events impact a wide variety of tree functions including reduced photosynthesis, increased photooxidative stress, leaves abscise, a decreased growth rate of remaining leaves and decreased growth of the whole tree (Teskey et al. 2015). Although trees are more resilient to heat stress (Teuling et al. 2010), it has been observed that different types of forest (e.g. needleleaf vs broadleaf) respond differently to drought and heat waves (Babst et al. 2012). For example, in the Turkish Anatolian forests NPP generally decreased during drought and heat waves events between 2000 and 2010, however, in a few other regions, NPP of needle leaf forests increased (Erşahin et al. 2016). However, forests may become less resilient to heat stress in future due to the long recovery period required to replace lost biomass and the projected increased frequency of heat and drought events (Frank et al. 2015; McDowell and Allen 2015; Johnstone et al. 2016; Stevens-Rumann et al. 2018).

Heat extremes also have an impact on fire. While ignition is largely related to human activities, the inter-annual variability of fire spread and frequency responds to large-scale climate fluctuations (Fernandes et al. 2011; Gutiérrez-Velez et al. 2014; Fanin and Van Der Werf 2017). Droughts have clear impact on fire occurrence (Davidson et al. 2012), although anomalously active fire seasons also occur during non-drought years, for example in Indonesia and in the Amazon (Gaveau et al. 2014; Brando et al. 2014). High temperatures increases the risk of fire through increase evapotranspiration rates that lead to greater soil and vegetation water depletion (Abatzoglou and Williams 2016; Fernandes et al. 2017; Aldersley et al. 2011). Even though humid tropical forest landscapes typically do not burn, the expansion of agriculture and deforestation into these landscapes make them vulnerable to drought-driven fires (Brando et al. 2014; Davidson et al. 2012). Seasonal fire anomalies are currently driven through seasonal to decadal fluctuations in rainfall (Fernandes et al. 2011), however, temperature is expected to become a more important factor than rainfall deficit as regional temperatures rise (Fernandes et al. 2017). In temperate and boreal regions fire seasons are lengthening and this trend is projected to continue in a warmer world (Flannigan et al. 2009; Williams and Abatzoglou 2016). However, future trends in future fire frequency, area and intensity especially at the regional scale are difficult to determine as there are complex and non-linear interactions between climate processes, (e.g., blocking highs) climate variables (e.g. temperature and humidity), atmospheric $CO_2$, fuels and human behaviour (Box 2.1).

Gross primary production (GPP) and soil respiration (Rs) form the first and second largest carbon fluxes from terrestrial ecosystems to the atmosphere in the global carbon cycle (Beer et al. 2010; Bond-Lamberty and Thomson 2010). Heat extremes impact the carbon cycle through altering these and change ecosystem-atmosphere $CO_2$ fluxes and the ecosystem carbon balance. Compound heat and drought events result in the strongest carbon sink reduction compared to single-factor extremes as gross primary production as (GPP) is strongly reduced and ecosystem respiration less so (Reichstein et al. 2013; Von Buttlar et al. 2018), although in forest biomes GPP may increase temporarily as a result of increased insolation and photosynthetic activity as was seen during the 2015/2016 ENSO related drought over Amazonia (Zhu et al. 2018). Longer extreme events (heat wave or drought or both) result in a greater reduction in carbon sequestration and may also reverse long-term carbon sinks (Ciais et al. 2005; Phillips et al. 2009; Reichstein et al. 2013; Wolf et al. 2016b; Ummenhofer and Meehl 2017; Von Buttlar et al. 2018). Furthermore, extreme heat events may impact the carbon cycle beyond the lifetime of the event. These lagged effects can slow down the carbon cycle, for example if reduced vegetation productivity and/or wide spread mortality after an extreme drought are not compensated by regeneration or accelerate it e.g. if productive tree and shrub seedlings cause rapid regrowth after windthrow or fire (Frank et al. 2015). Projected changes in the frequency and intensity of extreme temperatures (and drought) as a result of climate change are expected to result in decreased carbon sequestration by ecosystems, particularly in forests that due to their large carbon pools and fluxes, potentially large lagged impacts and long recovery times to regain lost stocks (Frank et al. 2015)(Section 2.4) .

#### *2.3.5.3 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. 2012; Trenberth 2012; Westra et al. 2013; Guhathakurta et al. 2017; Taylor et al. 2017; Thompson et al. 2017; Zilli et al. 2017) and the observed shift in the trend distribution for precipitation extremes is more distinct than for annual mean precipitation (Fischer and Knutti 2014). 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 as a result of the warming climate (Lehmann et al. 2015) and the global land fraction experiencing more intense precipitation events is larger than expected from internal variability (Fischer et al. 2013). A number of studies have attributed historical extreme rainfall events to human influence (Min et al. 2011; Pall et al. 2011; Sippel and Otto 2014; Trenberth et al. 2015) and the largest fraction of anthropogenic influence is evident in the most rare and extreme events (Fischer and Knutti 2014), however, the evidence for human influence on the probability of observed extreme precipitation events and storms is less robust than for temperature extremes (Stott 2016).

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 extreme precipitation events (Berg et al. 2013; Pall et al. 2007; Christensen and Christensen 2003; Wu et al. 2013; Thompson et al. 2017; Taylor et al. 2017; Zilli et al. 2017; Guhathakurta et al. 2017). Furthermore, changes to the dynamics of the atmospheric amplify or weaken future precipitation extremes at the regional scale, with higher intensification rates over the tropics and lower rates over the drier subtropics (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. 2012; Stott et al. 2016; Mohan and Rajeevan 2017; Prein et al. 2017). However, many climate models underestimate observed increased trends in heavy precipitation suggesting a substantially stronger intensification of future heavy rainfall than the multimodel mean (Min et al. 2011; Borodina et al. 2017). Furthermore, the response of convective precipitation extremes to warming remains uncertain because climate models are unable to explicitly simulate sub-grid scale convection, such as mesoscale convective systems, 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; Chawla et al. 2018; Kooperman et al. 2018; Maher et al. 2018; Rowell and Chadwick 2018), or capture the interannual variability in tropical precipitation extremes when compared to observations (Allan et al. 2010). Additionally, there is *low confidence* in the detection of long-term observed and projected seasonal and daily trends in extreme snowfall as a result of a relatively narrow range of temperatures below the rain–snow transition at which extreme snowfall can occur that is relatively insensitive to climate warming and subsequent large interdecadal variability (Kunkel et al. 2013; O'Gorman 2014, 2015).

#### *2.3.5.4 Impacts of precipitation extremes on land types and functioning*

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

In agricultural systems heavy precipitation inundation can delay planting, increases soil compaction, and causes crop losses through anoxia and root diseases (Posthumus et al. 2009) and in tropical regions flooding associated with tropical cyclones can lead to crop destruction and failure. In some cases flooding can affect yield more than drought, particularly in tropical regions (e.g., India) and in some mid/high latitude regions such as China and parts of France (Zampieri et al. 2017). Flooding can be beneficial in drylands as the floodwaters infiltrate and recharge alluvial aquifers along ephemeral river pathways, extending water availability to dry seasons and drought years and support riparian systems and human communities (Kundzewicz and Germany 2012; Guan et al. 2015). Waterlogging of croplands and soil erosion also negatively affect farm operations and block important transport routes (Vogel and Meyer 2018; Kundzewicz and Germany 2012). However, globally, the impact of extreme rainfall on agriculture is less than that of temperature extremes (Lesk et al. 2016).

Although many soils on floodplains regularly suffer from inundation, the projected increase in the magnitude of flood events means that new areas with no recent history of flooding are now becoming severely affected (Yellen et al. 2014). Surface flooding and associated soil saturation causes considerable losses in soil quality and plant productivity and induces changes in nutrient cycling with increased potential for nutrient loss, meso- and macro-faunal abundance, stimulates microbial growth and microbial community composition, negatively impacts redox and increases greenhouse gas emissions (Bossio and Scow 1998; Niu et al. 2014; Sánchez-Rodríguez et al. 2017; Barnes et al. 2018). The impact of flooding on soil quality is influenced by management systems that may mitigate or exacerbate the impact. Although soils tend to recover quickly after floodwater removal, the impact of repeated extreme flood events over longer timescales on soil quality and function is unclear.

Flooding in ecosystems may be beneficial, as a flood pulse bring nutrients to downstream regions or detrimental through erosion or permanent habitat loss (Kundzewicz et al. 2014). The effect of flooding on forests is not well studied although riparian forests can be damaged through flooding (Kramer et al. 2008). However, flooding is an important natural process in riparian forests and increased flooding may be of benefit in forests where upstream water demand has lowered stream flow, but this is difficult to assess (Pawson et al. 2013). Some grassland species under heavy rainfall and soil saturated conditions responded negatively with decreased reproductive biomass and germination rates (Gellesch et al. 2017), however overall productivity in grasslands remains constant in response to heavy rainfall (Grant et al. 2014).

Soil $CO_2$ fluxes and $CO_2$ uptake by plants within ecosystems are altered in response to extreme rainfall and therefore result in changes in ecosystem carbon cycling (Fay et al. 2008; Frank et al. 2015). Extreme rainfall and flooding limits oxygen in soil which may suppress the activities of soil microbes and plant roots and lower soil respiration and therefore carbon cycling (Knapp et al. 2008; Rich and Watt 2013; Philben et al. 2015). However, the impact of extreme rainfall on carbon fluxes in different biomes differs. For example, extreme rainfall in mesic biomes reduces soil $CO_2$ flux to the atmosphere and GPP whereas in xeric biomes the opposite is true, largely as a result of increased soil water availability (Knapp and Smith 2001; Heisler and Knapp 2008; Heisler-White et al. 2009; Xu and Wang 2016; Liu et al. 2017; Connor and Hawkes 2018). Furthermore, changes in heavy rainfall in the dry season may have a greater impact than in the colder season, long intra-seasonal dry periods that are interspersed with a few heavy rainfall events can result in increased productivity due to increased soil water availability (Zeppel et al. 2014; Heisler-White et al. 2009; Liu et al. 2017).

## 2.4 GHG fluxes between land and atmosphere

The Paris Agreement requires credible estimates of anthropogenic fluxes of greenhouse gasses (GHGs), both “emissions” and “removals” (Fuglestvedt et al. 2018). Land is simultaneously a source and sink for several greenhouse gasses. Moreover, both natural and anthropogenic processes drive fluxes of GHGs, making it difficult to separate “anthropogenic” and “non-anthropogenic” emissions and removals. The IPCC divided the processes responsible for fluxes of carbon dioxide from land into three categories (IPCC 2010): (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, $CO_2$ fertilisation, N deposition; and (3) *natural* climate variability and natural disturbances (e.g. wildfires, windrow, disease). The IPCC (2010) noted that it was impossible with any direct observation to separate anthropogenic from non-anthopogenic fluxes 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 Paris Agreement includes an Enhanced Transparency Framework, to track countries’ progress towards achieving their individual targets (i.e., the Nationally Determined Contributions, 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. It is expected that the global stocktake will compare country reports of Greenhouse Gas Inventories (GHGIs) submitted to the United Nations Framework Convention on Climate Change (UNFCCC) with modelled mitigation pathways (Grassi et al. 2018). This expectation implies a need to ensure consistency between different approaches or, if they are not consistent, to assess why and if they can be reconciled (Grassi et al. 2018; Fuglestvedt et al. 2018). Here we assess recent estimates of anthropogenic GHG emissions and removals from the scientific literature (included in AR5) and from country GHGIs (not included in AR5).

The major GHGs exchanged between land and the atmosphere discussed in this chapter are carbon dioxide ($CO_2$, Section 2.4.1), methane ($CH_4$, Section 2.4.2) and nitrous oxide ($N_2O$, Section 2.4.3). Emissions from Agriculture, Forestry, and Land Use (AFOLU) were responsible for approximately 25% of anthropogenic GHG emissions over the period 2000-2010 (AR5) (Smith et al. 2013a; Ciais et al. 2013). In the sub-sections below, we update the estimates for each gas, finding that AFOLU, as estimated by global carbon models, contributed 26% of anthropogenic GHG emissions. We also describe new insight into the differences between global carbon models and GHGIs in estimating the “anthropogenic” sink

### 2.4.1 Carbon dioxide

#### *2.4.1.1 Land in the Global Carbon Budget-the total net flux of $CO_2$ between land and atmosphere*

The continuous exchange of $CO_2$ between land and the atmosphere as part of the natural global carbon cycle includes an annual gross removal of about 440 $GtCO_2\ yr^{-1}$ (120 $GtC\ yr^{-1}$) due to photosynthesis, and annual gross emissions of nearly the same magnitude to the atmosphere due to respiration by plants, animals, and microbes (Ciais et al. 2013). Many processes lead to changes in these gross fluxes. The net effects of all anthropogenic and non-anthropogenic processes on managed and unmanaged land is a net removal of $CO_2$ from the atmosphere (*high agreement, robust evidence*). This total net land-atmosphere removal is estimated to have been 5.1 to 8.4 $GtCO_2$ $yr^{-1}$ 2007 to 2016 across a range of approaches based on atmospheric flux measurements and satellite estimates of $CO_2$ concentration used with inverse modelling techniques (Le Quéré et al. 2018; van der Laan-Luijkx et al.,2017; Rödenbeck, 2005; Chevallier et al., 2005) (See Box 2.1). While inverse modelling can separate the net land flux from fossil fuel flux and ocean fluxes, they are unable to further separate the land $CO_2$ flux into anthropogenic versus non-antrhopogenic fluxes.

The Global Carbon Budget is evaluated each year by the Global Carbon Project (Le Quéré et al. 2018) and models are used to calculate, separately, fluxes due to AFOLU and due to environmental change (Section 2.4.1.2; Table 2.1, Box 2.1). The sum of the modelled estimates gives a total net land-atmosphere flux of 6.3_ ± 2.6 $GtCO_2$ $yr^{-1}$ over the period 2007-2016. Although the land has been a net sink for $CO_2$ since around the middle of last century, it was a net source to the atmosphere before that time (Le Quéré et al. 2018).

At the present time, vegetation stores about 450 Gt of carbon (range 380-536) (Erb et al. 2018), while soils (to a depth of 2 meters) store about 3195 GtC (Sanderman et al. 2017). Prior to large scale AFOLU activities, potential vegetation is estimated to have stored about 916 (range 771-1107) PgC under current climate conditions (Erb et al. 2018), while soils stored about 3311 GtC (Sanderman et al. 2017). The difference between potential and current stocks highlights the massive effect of direct and indirect anthropogenic changes in reducing biomass (466 GtC) and soil carbon (116 GtC) stocks (total loss = 582 GtC). Deforestation and other changes in land cover (e.g., conversion of forest to cropland) were responsible for 53-58% of the difference in biomass stocks (Erb et al. 2018). Land management, or stock changes induced by land use within the same type of land cover, contributed 42%-47%. Previous estimates of the loss of carbon from land use, including both vegetation and soils, have varied between 158 and 545 PgC (Houghton et al. 2012; Pongratz et al. 2010; Kaplan et al. 2011).

**Table 2.1 Perturbation of the global carbon cycle caused by anthropogenic activities ($GtCO_2$ $yr^{-1}$) (Le Quéré et al. 2018)**

| | Mean ($GtC\,yr^{-1}$) | | | | | | |
|---|---|---|---|---|---|---|---|
| | 1960–1969 | 1970–1979 | 1980–1989 | 1990–1999 | 2000–2009 | 2007–2016 | 2016 |
| Emissions | | | | | | | |
| Fossil fuels and industry ($E_{FF}$) | 3.1 ± 0.2 | 4.7 ± 0.2 | 5.5 ± 0.3 | 6.3 ± 0.3 | 7.8 ± 0.4 | 9.4 ± 0.5 | 9.9 ± 0.5 |
| Land-use change emissions ($E_{LUC}$) | 1.4 ± 0.7 | 1.1 ± 0.7 | 1.2 ± 0.7 | 1.3 ± 0.7 | 1.2 ± 0.7 | 1.3 ± 0.7 | 1.3 ± 0.7 |
| Partitioning | | | | | | | |
| Growth rate in atmospheric $CO_2$ concentration ($G_{ATM}$) | 1.7 ± 0.1 | 2.8 ± 0.1 | 3.4 ± 0.1 | 3.1 ± 0.1 | 4.0 ± 0.1 | 4.7 ± 0.1 | 6.0 ± 0.2 |
| Ocean sink ($S_{OCEAN}$) | 1.0 ± 0.5 | 1.3 ± 0.5 | 1.7 ± 0.5 | 1.9 ± 0.5 | 2.1 ± 0.5 | 2.4 ± 0.5 | 2.6 ± 0.5 |
| Terrestrial sink ($S_{LAND}$) | 1.4 ± 0.7 | 2.4 ± 0.6 | 2.0 ± 0.6 | 2.5 ± 0.5 | 2.9 ± 0.8 | 3.0 ± 0.8 | 2.7 ± 1.0 |
| Budget imbalance | | | | | | | |
| $B_{IM} = E_{FF} + E_{LUC} - (G_{ATM} + S_{OCEAN} + S_{LAND})$ | (0.4) | (−0.6) | (−0.4) | (0.1) | (0.0) | (0.6) | (−0.2) |

### *2.4.1.2 AFOLU and non-AFOLU land fluxes[1]*

Land has been a net source of carbon emissions due to AFOLU activities over recent decades *(high agreement*, *robust evidence)* although there is a wide range among estimates. A variety of different definitions, methods and approaches are used for estimating anthropogenic fluxes of carbon from land due to AFOLU, each relevant to the different purposes, methods, data and scope for which it was developed (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 2.1 and Figure 2.8 and Figure 2.9).

[1] FOOTNOTE: The term "land-use change", as used in the scientific literature, is conceptually similar to LULUCF (Land Use, Land-Use Change and Forestry) in that it combines impacts of changes in cover (forest, cropland) and some management (e.g. wood harvest) on $CO_2$ flux. In AR5 it was referred to as the "FOLU" part of AFOLU. Here we use the term AFOLU across the whole special report, although noting that it does not include non-$CO_2$ GHG fluxes from agriculture, which are covered in 2.4.2 and 2.4.3.

*Variability among estimates*

The DGVMs used spatially explicit, harmonised land-use change data (LUH2) (Hurtt et al. 2017) based on HYDE 3.2. The Houghton bookkeeping approach (Houghton and Nassikas 2017) used land cover-change data from the FAO Forest Resource Assessment (FAO 2015) and FAOSTAT, and estimates of peat burning emissions from the Global Fire Emissions Database (GFED version 4, (Randerson et al. 2015). Houghton and Nassikas (2017) did not include shifting cultivation, but they did include Indonesian and Malaysian peat burning and drainage (from (Hooijer et al. 2010)).

Satellite-based estimates of $CO_2$ emissions from loss of tropical forests during 2000-2010 corroborate the modelled emission but are quite variable: 4.8 $GtCO_2$ $yr^{-1}$ (Tyukavina et al. 2015), 3.0 $GtCO_2$ $yr^{-1}$ (Harris et al. 2015), 3.2 $GtCO_2$ $yr^{-1}$ (Achard et al. 2014) and 1.6 $GtCO_2$ $yr^{-1}$ (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 adjusted their findings for bias. Three of the estimates (except (Baccini et al. 2017) considered losses in forest area and ignored degradation and regrowth of forests. Baccini et al. (2017), on the other hand, included both losses and gains in forest area and losses and gains of carbon within forests (i.e., forest degradation and growth). Together, these processes yielded a lower total loss, presumably because of forest growth. Some of the growth in carbon stocks results from recovery of forests following harvest of wood or agricultural abandonment (i.e., direct anthropogenic effect) (Houghton and Nassikas 2017; Hyrynsalmi et al. 2017), and some is thought to result from $CO_2$ fertilisation (an indirect effect) (Schimel et al. 2015). The four 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, some of the carbon in trees is not released immediately to the atmosphere but transferred to dead and downed vegetation or wood products. Both bookkeeping models and DGVMs account for the delayed emissions in growth and decomposition.

In addition to differences in land-cover data sets between models and satellites, there are many other methodological reasons for differences (See Box 2.1)(Gasser and Ciais 2013; Pongratz et al. 2014; Vermeulen et al. 2012b; Tubiello et al. 2015; Houghton et al. 2012). 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 (MgC/ha), 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 fairly recently (since AR5) included forest management processes such as tree harvesting and land clearing for shifting cultivation, leading to larger estiamtes of $CO_2$ emissions than when these processes are not considered (Arneth et al. 2017; Erb et al. 2018; Luyssaert et al. 2014).. Grazing management has likewise been found to have large effects (Sanderman et al. 2017), not included in most DGVMs (Pugh et al. 2015; Pongratz, 2018). Other unexplained discrepencies remain in some regions.

The wide range of estimates of net fluxes of $CO_2$ due to AFOLU may overestimate uncertainty, however, because taking account of the different approaches sometimes identifies important processes and can help with transparency and credibility in monitoring, reporting and verifying GHG fluxes under the UNFCCC, e.g. (Grassi et al. 2018).

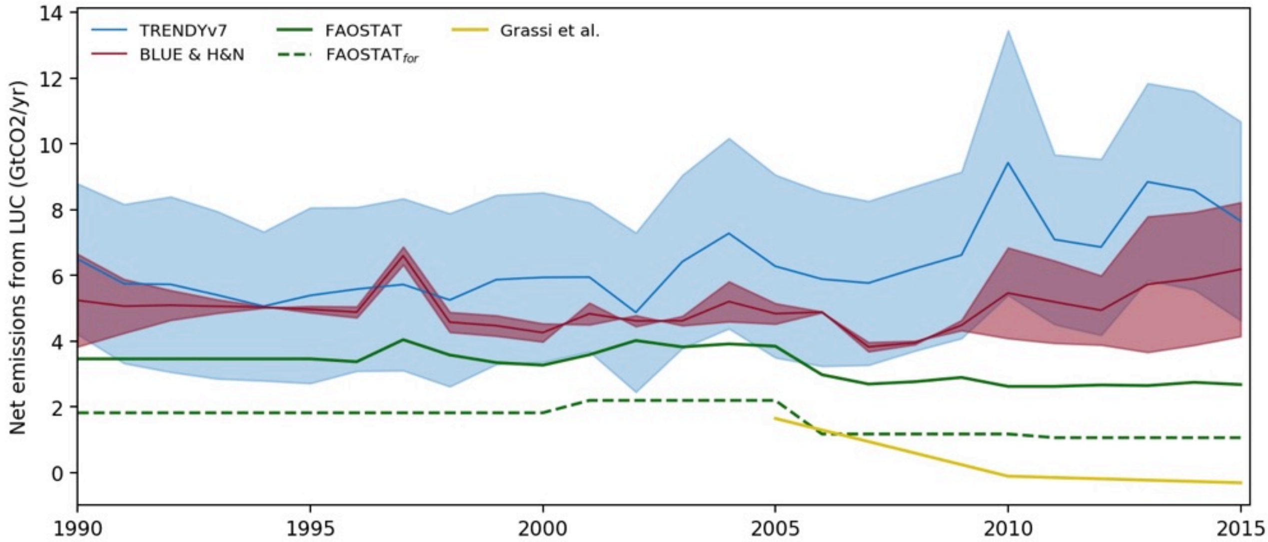


**Figure 2.7 Global net CO2 emissions due to AFOLU (Agriculture, Forestery and Other Land Use) from a range of different approaches (in GtCO2/yr) Red line: the mean and range across two bookkeeping models (Houghton and Nassikas 2017; Hansis et al. 2015) used for the "land use change" flux estimated in the Global Carbon Budget (Le Quéré et al. 2018). Blue line: the mean from Dynamic Global Vegetation Models run with the same driving data for the Global Carbon Budget (Le Quéré et al. 2018) with the pale blue shading showing the 1 standard deviation range. Green line: FAO (2015) (downloaded from FAOSTAT website), the dashed line is primarily forest-related emissions, while the solid green line also includes emission from peat fires and peat draining. Yellow line: Greenhouse Gas Inventories GHGI based on country reported data to UNFCCC compiled by (Grassi et al. 2018). Data are shown only from 2005 because reporting in many developing countries became more consistent/reliable after this date. For more details on methods see Box 2.1.**

*Land in the Global Carbon Budget*

The 2017 Global Carbon Budget (Le Quéré et al. 2018) partitions the net land-atmosphere flux into two terms: "land use change" (considered anthropogenic $CO_2$ fluxes due to AFOLU) and the "land sink" due to indirect anthropogenic effects. The modelled "land use change" (AFOLU) flux was a net emission of 4.9 ± 3.0 $GtCO_2$ $yr^{-1}$ for 2007-2016, approximately 12% of total anthropogenic $CO_2$ emissions ((Le Quéré et al. 2018), Table 2.1). This net flux is 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 (not all models include all activities). The land-use change flux in the Global Carbon Budget is the mean of two bookkeeping models (Houghton and Nassikas 2017; Hansis et al. 2015), but is similar to the mean across a range of Dynamic Global Vegetation Models (DGVMs) (Le Quéré et al. 2018, Box 2.1, Figure 2.7). Net CO2 emissions from AFOLU have stayed relatively constant since 1900. AFOLU emissions were the dominant anthropogenic source until around the middle of the last century when fossil fuel emissions became more dominant (Le Quéré et al. 2018).

According to Global Carbon Budget estimates (Le Quéré et al. 2018), just over half of total $CO_2$ emissions (LULUCF and fossil fuels) were taken up by ocean and land removals (Table 2.1) *(robust evidence, high agreement)*. The land sink is driven largely by the indirect effects of environmental change (climate, $CO_2$, N deposition) on both managed and unmanaged land. As described in Section 2.2, rising $CO_2$ concentrations have a fertilising effect on land, while climate has a mixture of effects; e.g., rising temperature increases respiration rates and may enhance or reduce photosynthesis depending on location and season, while longer growing seasons allow for higher photosynthesis. The net "land sink" due to the indirect effects of environmental change has increased steadily since 1900 and was -11.2 ± 3.0 $GtCO_2$ $yr^{-1}$ during the period 2007 to 2016, absorbing 22% of global anthropogenic emissions.

The land sink may have slowed the rise in global land-surface air temperature by 0.09 ± 0.02°C since 1982 (*medium confidence*) (Zeng et al. 2017). The rate of $CO_2$ removal by land accelerated from -0.007 ± 0.065

GtC $yr^{-2}$ during the warming period (1982 to 1998) to -0.119 ± 0.071 GtC $yr^{-2}$ during the warming hiatus (1998-2012), perhaps from lower respiration rates during the warming hiatus (Ballantyne et al. 2017). On the other hand, the lower rate of growth in atmospheric $CO_2$ during the warming hiatus may have resulted, not from lower rates of respiration, but from declining emissions from LULUCF (lower rates of tropical deforestation and increased forest growth in northern mid-latitudes (Piao et al. 2018). Changes in the growth rate of atmospheric CO2, by themselves, do not identify the processes responsible.

While year to year variability in the indirect land sink is high in response to climate variability, DGVM fluxes on decadal time scales are far more influenced by $CO_2$ fertilisation. A DGVM intercomparison (Sitch et al. 2015) for 1990 to 2009 found that $CO_2$ alone contributed a mean global removal of -2.875 ± 1.003 GtC $yr^{-1}$ (trend -0.121 ± 0.055GtC $yr^{-2}$) while climate alone contributed a source of 0.497 ± 0.523 GtC $yr^{-1}$ (trend 0.039 ± 00.022PgC $yr^{-2}$) giving a net indirect effect of -2.378 ± 00.721 GtC $yr^{-1}$ (trend -0.055 ± 00.03PgC $yr^{-2}$). Data from forest inventories around the world corroborate the modelled land sink (Pan et al., 2011). A recent analysis of changes in aboveground biomass throughout the tropics suggests that the residual sink is distributed between the extra-tropics and tropics in the ratio of 2:1 (Houghton and Nassikas 2017). . That estimate is uncertain because it did not account for changes in belowground carbon (soils).

The land sink was referred to in AR5 as the “residual terrestrial flux,” as it was not estimated directly, but calculated by difference from the other directly estimated fluxes in the budget (Table 2.1). In the 2017 budget, the land sink term was instead estimated directly by DGVMs, leaving a budget imbalance of 2.2 Gt $CO_2$ $yr^{-1}$ (sources overestimated or sinks underestimated). The budget imbalance may result from variations in oceanic uptake or from land processes not included in models. For example, a decline in the global area burned by fires each year (Andela et al. 2017) (boreal forests represent an exception to this decline (Kelly et al., 2013)) accounts for an estimated net sink of 0.5 Gt $CO_2$ $yr^{-1}$ (Arora and Melton 2018). There is also an estimated net carbon sink of about the same magnitude in anaerobic sediments of aquatic environments as a result of soil erosion (from agricultural lands) and redeposition in anoxic environments (WANG et al. 2018) .

*Regional differences*

Figure 2.8 shows regional differences in emissions due to AFOLU, and the difference between 1990-1999 and 2006 to 2015. Recent increases in deforestation rates in some tropical countries have been balanced by increases in forest area in India, China, the USA and Europe (FAO-FRA 2015). The trend in emissions from AFOLU since the 1990s is *uncertain* because some data suggest a declining rate of deforestation (FAO 2015), while other data suggest an increasing rate (Hansen et al. 2013; Kim et al. 2014). The disagreement results in part from differences in the definition of forest and approaches to estimating deforestation. The FAO defines deforestation as the conversion of forest to another land cover (FAO, UNFCCC), while the measurement of forest loss by satellite may include wood harvests (forests remaining forests) and natural disturbances that are not directly caused by anthropogenic activity (e.g., forest mortality from droughts and fires). However, the difference between satellite and ground inventory data may reveal real differences. For example, recent drought-induced fires in the Amazon have increased the emissions from wildfires at the same time that emissions from deforestation have declined (Aragão et al. 2018a).

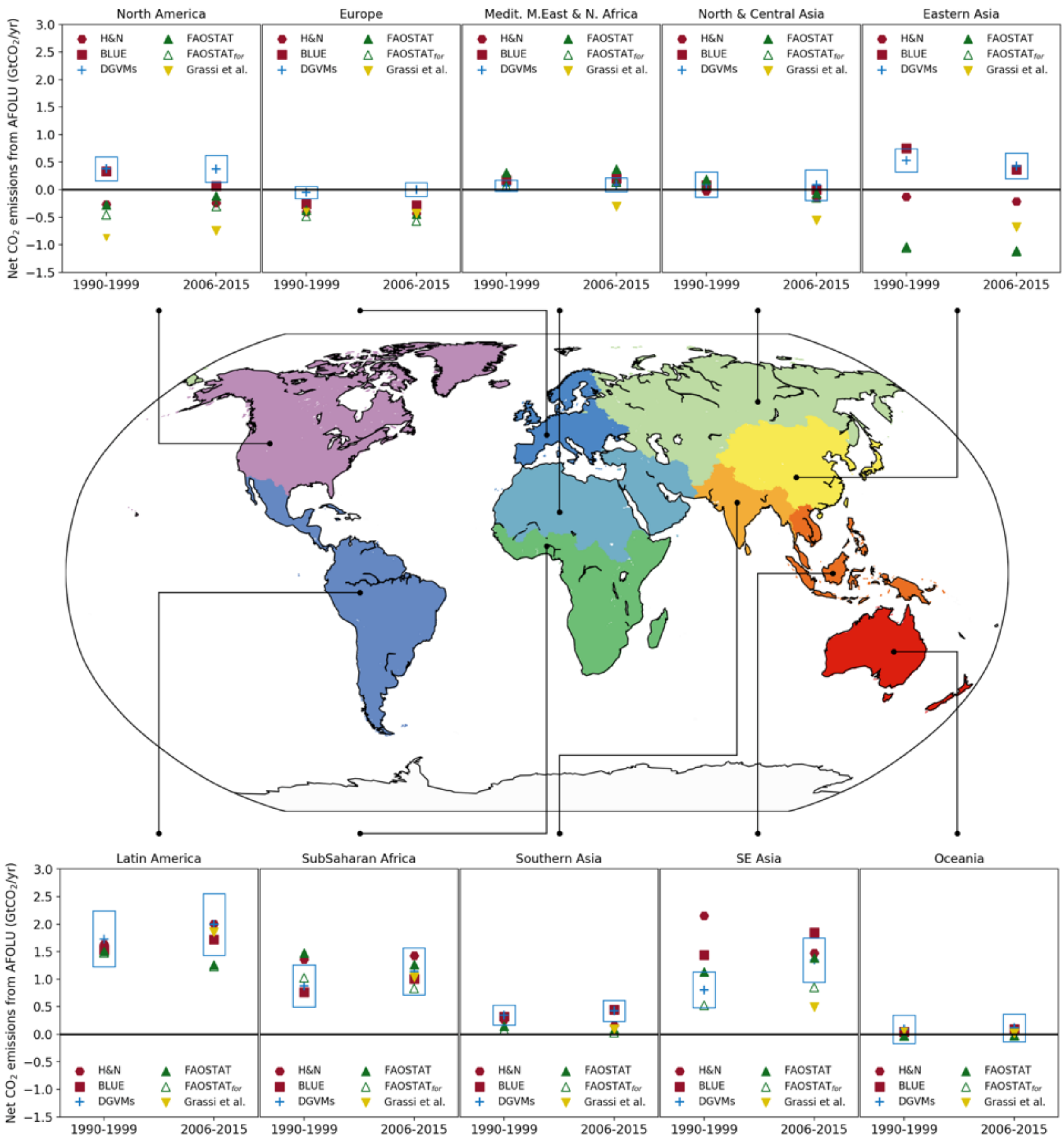


**Figure 2.8 Regional trends in net CO2 emissions due to AFOLU (Agriculture, Forestery and Other Land Use) from a range of different approaches (in GtCO2 yr-1). Red symbols: bookkeeping models (trianle - (Houghton and Nassikas 2017; Hansis et al. 2015). Blue cross: the mean from Dynamic Global Vegetation Models run the Global Carbon Budget** (Le Quéré et al. 2018) **with the box showing the 1 standard deviation range. Green circles: FAO (2015) (downloaded from FAOSTAT website), the open circle is primarily forest-related emissions, while the closed circle includes emission from peat fires and peat drainage. Yellow circle: Greenhouse Gas Inventories GHGI based on country reported data to UNFCCC compiled by (Grassi et al. 2018) – data for developing countries is only shown for 2006-2015 because reporting in many developing countries became more consistent/reliable after 2015. For more details on methods see Box 2.1**

*Nationally Reported GHGI values versus Global Model Estimates*

Aggregated country GHGI estimates are lower globally than global model-based estimates (Figure 2.7, Grassi et al. 2017; Grassi et al. 2018) and the same is true in many regions (Figure 2.8). Fluxes reported to the UNFCCC in country GHGIs were found to be around 4.3 $GtCO_2$ $yr^{-1}$ lower (Grassi et al. 2018) than the estimate from the bookkeeping model (Houghton et al. 2012) used in the carbon budget for AR5 (Ciais et al. 2013; Tanaka and O'Neill 2018). The conceptual differences between GHGIs, IPCC AR5 (bookkeeping model), and the latest approach of the Global Carbon Project in estimating the "anthropogenic land flux" are illustrated in Figure 2.9. GHGIs include all fluxes occurring on "managed lands" (i.e., direct, indirect and natural effects) as AFOLU. The bookkeeeping models include only the direct anthropogenic effects of AFOLU, without explicitly modelling indirect or natural effects. The DGVMs, by differencing runs with and without land use change, include both direct and indirect effects on land experiencing AFOLU as part of the "land use change flux". Indirect effects on other land are part of the "land sink".

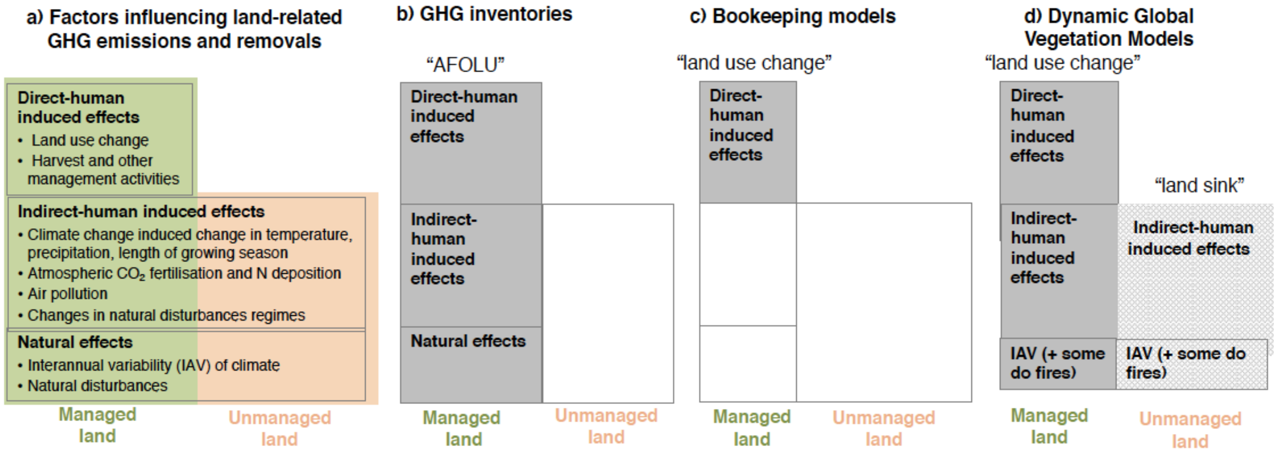


**Figure 2.9 Summary of the main conceptual differences between GHG Inventories and global models in considering what is the "anthropogenic land CO2 flux". Adapted from Grassi et al. (2018): a) Effects of key processes on the land flux as defined by** (IPCC 2010) **including where these effects occur (in unmanaged/primary lands, vs. managed/secondary lands); How these effects are captured in (b) GHG Inventories reported by countries to the UNFCCC that report all fluxes in "managed land" but do not report unmanaged land; '(c) bookkeeping models that do not explicitly model the effects of environmental change, although some is implicitly captured in data on carbon densities and growth and decay rates; and (d) Dynamic Global Vegetation Models (DGVMs) that include the effects of environmental change on all lands, and run the models with and without land use change to diagnose "land use change", the "land sink" is then conceptually assumed to be a natural response of land to the anthropogenic perturbation of environmental change, the area of land considered subject to management activities is less than that considered "managed land" in GHG inventories, see Figure 2.11.**

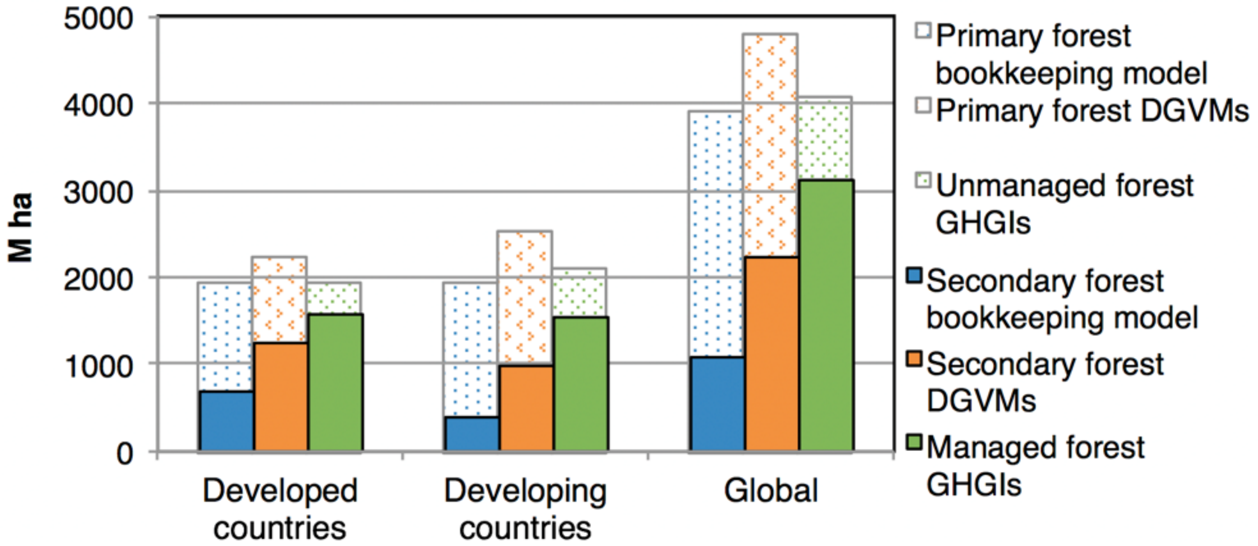


**Figure 2.10 Differences in areas of "managed" and "unmanaged" forests between GHG Inventories and global models. Figure from (Grassi et al. 2018) based on data from the bookkeeping model of (Houghton and Nassikas 2017), Dynamic Global Vegetation Models (DGVMs) included in the Global Carbon Budget (Le Quéré et al. 2018) and country data submitted to UNFCCC compiled by Grassi et al. (2018). The analysis does not include all developing countries, but covers about 85% of the FAO-FRA's global "secondary forest" area (FAO-FRA 2015)**

Flux estimates from global models are similar to those from GHGIs for deforestation and afforestation/reforestation, as they are based on data largely in agreement. Global models use land-cover change from (FAO 2015) , which is broadly consistent with data reported in country inventories. However, global models and GHGIs have different estimates of fluxes from "managed" forests, particularly in northern extra-tropical regions (Figure 2.8). The bookkeeping models (Houghton et al. 2012; Hansis et al. 2015) and some DGVMs (Sitch et al. 2015; Le Quéré et al. 2016) include land management (wood harvest and regrowth) explicitly. In contrast, the GHGIs' "managed land" concept is broader and includes activities related to the social and ecological functions of land. The "managed land" areas considered by GHGIs are generally much larger than the lands experiencing harvests in global models (Figure 2.10). Grassi et al. (2018) used post processing of the DGVM results from the Global Carbon Project to estimate the indirect effects on areas of "secondary" versus "managed" forests. The indirect effects on the larger area of managed forests accounted for 3.3 $GtCO_2$ $yr^{-1}$, or 75% of the discrepancy between GHGIs and the bookkeeping model.

Further reconciliation of the differences between estimates would enable a more credible Global Stocktake (Grassi et al. 2018; Fuglestvedt et al. 2018). And further reconciliation would by possible if GHGIs were to provide more transparent and complete information on managed forests (including maps, harvested area, harvest cycle, forest age and if/how indirect and natural effects are included). Likewise, since the global models use data submitted by countries to FAO, it would enhance transparency if countries reported consistently to both the UNFCCC and FAO, which currently is not always the case (Frederici et al., 2017). Finally, there are opportunities for the global modeling community, including Integrated Assessment Models, to design future models and model experiments to increase their comparability with historical GHGIs, and thus their relevance in the context of the Paris Agreement, including more details on forest management and disaggregated results for fluxes from managed lands.

The modelled AFOLU flux of 4.9 ± 3.0 $GtCO_2$ $yr^{-1}$ over the period 2007 to 2016 represents a net value. It consists of both gross emissions of $CO_2$ from deforestation, forest degradation, and the oxidation of wood products, as well as gross removals of carbon from carbon accumulation in forests and soils recovering from harvests and agricultural abandonment. Few studies report gross fluxes due to AFOLU, however. (Houghton and Nassikas 2017) estimated gross emissions to be as high as 20.2 Gt $CO_2$ $yr^{-1}$ (Figure 2.11). Even this may be an underestimate as the land-cover change data used from FAO (FAO 2015)is in itself a net change. If the areas of croplands are increasing in some parts of a country and decreasing in others, only the net change is reported to the FAO.

These gross fluxes are more informative for assessing the potential for mitigation than estimates of net fluxes, because the gross fluxes indicate the extent of individual activities. Gross emissions from rotational land use in the tropics are approximately 37% of total $CO_2$ emissions, rather than 10-12%, as suggested by net AFOLU emissions (Houghton and Nassikas 2018a). Further, gross removals of nearly the same magnitude would be expected to continue for decades if deforestation were to be eliminated.

Gross emissions and removals of $CO_2$ result from rotational uses of land, such as wood harvest and shifting cultivation. Evidence for rotational land uses comes indirectly from the FAO (FAOSTAT (FAO-FRA 2015): only about 2/3 of deforestation in the tropics is for increased agricultural areas (permanent croplands or pastures). Nearly a third of the lands deforested lead to increases in “other land” (not forest, cropland, or pasture, but perhaps shifting cultivation). The observation is too large and widespread among countries to be explained by error or other land uses (Houghton and Nassikas 2018a). Rates of net forest area loss given by the FAO-FRA (2015) are consistent with rates of deforestation observed with satellite data.

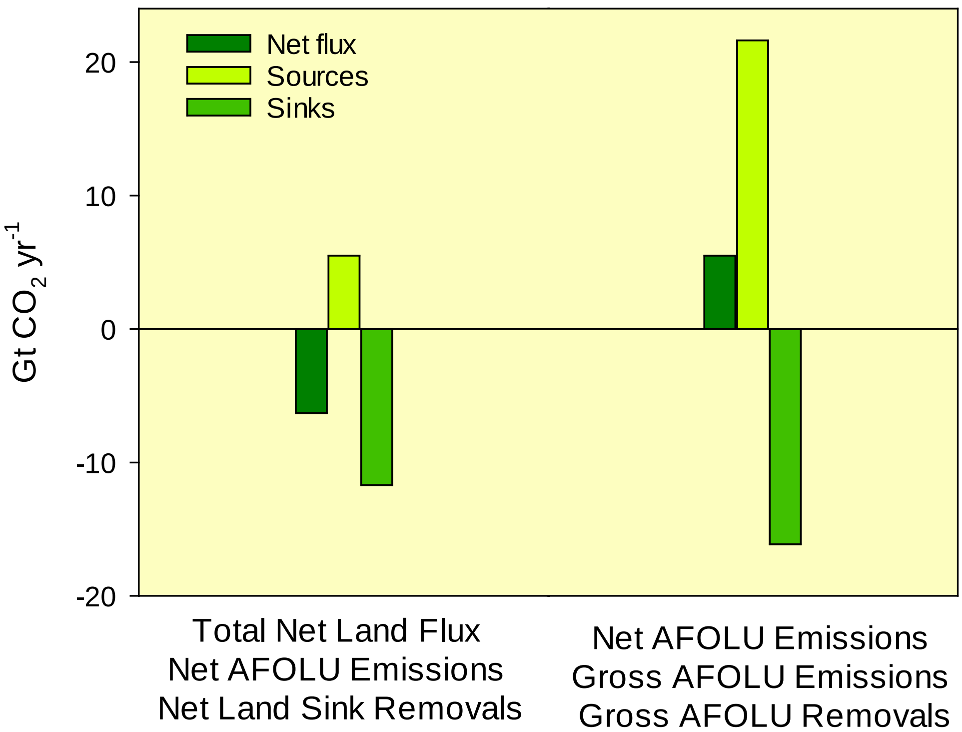


**Figure 2.11 Net fluxes, sources, sinks for AFOLU and for all lands. The bars for the left-hand group show the total net land flux and how this partitions into net AFOLU emissions and net AFOLU sink removals. The bars in the right-hand group shows how the net AFOLU emissions partition into gross sources and sinks. There are no global estimates to partition the net land sink into its gross components**

Estimates of gross emissions and removals for indirect and natural fluxes of carbon do not exist at the global level, although there are many studies that suggest increasing emissions of carbon from such effects, for example: temperature-induced increases in respiration rates (Bond-Lamberty et al. 2018); increased tree mortality (Berdanier and Clark 2016; McDowell et al. 2018); and thawing permafrost (Schuur et al. 2015). Nevertheless, the global carbon budget suggests that land and ocean sinks have actually *increased* over the last five decades in proportion to total $CO_2$ emissions (Le Quéré et al. 2018). There is little evidence for a reduction in the net removal. That means that large emissions from indirect-anthropogenic and natural effects must have been balanced by even larger removals (driven, as discussed in section 2.4.1.2) by $CO_2$ fertilization, climate change, N deposition, erosion and redeposition of soil carbon, and changes in natural disturbances). Gross emissions and gross removals must be larger than the global net removal, but only the net removal is known. The magnitude of the gross fluxes is unknown and unconstrained.

#### *2.4.1.3 Impact of climate change on future fluxes*

Climate change is expected to impact terrestrial biogeochemical cycles via an array of complex feedback mechanisms that will act to either enhance or decrease future $CO_2$ emissions (described in Section 2.2 and 2.3). The balance of these positive and negative feedbacks remains uncertain. Estimations from climate models included in AR5, CMIP5 and $C^4$MIP exhibit large differences for different carbon and nitrogen cycle feedbacks and how they change in a warming climate (Anav et al. 2013; Friedlingstein et al. 2006; Friedlingstein et al. 2014). The differences are in large part due to the uncertainty regarding how primary productivity will respond to environmental changes, with many of the models not even agreeing on the sign of change. Furthermore, many models do not include a nitrogen cycle, which limits the $CO_2$ fertilisation effect. These uncertainties are further exacerbated by the lack of observational constraints (Prentice et al. 2015a).

The $CO_2$ fertilisation effect is expected to increase $CO_2$ uptake, which in addition to a decrease in stomatal conductance may drive an increase in productivity and a consequent greening effect. However, given that plant biomass has fixed C:N ratios (although they vary by plant and soil type), the magnitude and persistence of the $CO_2$ fertilisation effect depends on the availability of mineral N and the ability of plants to acquire it. N limitation, particularly in natural ecosystems, is a key limiting factor on C storage capacity, which is not yet well represented in many Earth System models (Zaehle et al. 2015). Similarly, $CO_2$ fertilisation is likely to be limited by an upper limit to the efficiency of photosynthesis (Heimann and Reichstein, 2008). In contrast, increased temperature may enhance microbial decomposition and result in greater $CO_2$ release from soils (Heimann and Reichstein, 2008), as well as enhanced degradation of permafrost and wetland ecosystems. The shift in vegetation distribution due to climate change may also impact carbon and nitrogen cycles.

Analysis by Prentice et al. (2015b) that looked specifically at the land $CO_2$ feedbacks in the CMIP5 and $C^4$MIP models under a range of scenarios indicated that negative feedbacks associated with increased productivity would outweigh the positive feedbacks throughout the 21st century. Arneth et al. (2010) came to a similar conclusion, showing that the feedback range for $CO_2$ fertilisation was between -0.17 to -1.9 $Wm^{-2}$ compared to 0.1 to 0.9 $Wm^{-2}$ for the positive feedbacks. In both studies however, the inclusion of the nitrogen cycle reduces this $CO_2$ fertilisation effect significantly due to nutrient limitations. Arneth et al. (2010) showed that the negative feedbacks decreased to -0.4 to -0.8 $Wm^{-2}$, which for some models removed the net carbon loss completely. Similarly, when including the feedbacks associated with methane, fire and ozone, the total radiative forcing between the atmosphere and terrestrial biosphere was positive, ranging between 0.9 to 1.5 $Wm^{-2}$ $K^{-1}$ at the end of the 21st Century. Other studies suggest that the cumulative warming effect of methane ($CH_4$) and $N_2O$ emissions over the period 2001-2010 was a factor of two larger than the cooling effect that resulted from $CO_2$ fertilisation (Tian et al. 2016). If so, mitigation efforts should be as focused on reducing emissions of these non-$CO_2$ GHGs as on increasing carbon storage capacity. Nevertheless, studies highlight the uncertainty in carbon-cycle estimates from ESMs and the need for more realistic carbon and nutrient cycling in models.

It is uncertain whether warmer climates will lead to a greater or lesser storage of carbon on land. Vegetation holds more carbon in warm regions (the tropics), while soils hold more carbon in cool regions (temperate zone and boreal regions).

### 2.4.2 Methane

#### *2.4.2.1 Methods – CH4*

As for $CO_2$, several methods are applied to estimate global methane fluxes. Process models for wetlands and fire emissions are parameterised for local conditions which are then driven by global climate data or satellite observations of burned area. These data are complimented by emissions inventories of agricultural activities, energy production and use, and sector specific emission factors to provide yearly or periodic average emissions estimates. Many studies combine “top-down” atmospheric measurements and inversions with the DGVM “bottom-up” model and inventory estimates to look for consistency between the approaches for the different terms of $CH_4$ budgets. These approaches are not completely independent as bottom-up estimates are typically used in inversion modelling to describe “prior” spatial distributions of sources and sinks, which are then modified by the inverse model (Combal et al., 2003; Bergamaschi et al., 2013).

Hydroxyl radicals (OH) are an important part of atmospheric chemistry and they can be thought of as the detergent of the atmosphere because they react with many pollutants and contribute to removing them from the atmosphere. OH reacts with $CH_4$ in the first step toward full oxidation to $CO_2$. In global $CH_4$ budgets, the atmospheric OH sink is difficult to quantify because the radical has a lifetime on the order of 1 second and its distribution is controlled by different precursor species that have non-linear interactions (Taraborrelli et al., 2012; Prather et al., 2017). Results from the Atmospheric Chemistry and Climate Model Intercomparison Project (ACCMIP) (Voulgarakis et al. 2013) produced a series of bottom-up, time-slice experiments that estimated long-term changes in atmospheric composition. As most models do not produce year to year estimates of the OH variability, time-slice results are used in most $CH_4$ budgets. These bottom-up estimates can be adjusted at large scales using inversion models based on measurements of tracers such as methyl chloroform or chloromethanes that have known emissions and that are removed through reactions with OH (Kirschke et al. 2013).

#### *2.4.2.2 Atmospheric trends*

In 2016, the globally averaged atmospheric concentration of $CH_4$ was 1843 ± 1 ppbv (Figure 2.12a). Systematic measurements of atmospheric $CH_4$ concentrations began in the mid-1980s and trends show a steady increase between the mid-1980s and early-1990s, slower growth thereafter until 1999, a period of no growth between 1999 and 2006, followed by a resumption of growth in 2007 that continues. The growth rates show very high inter-annual variability with a negative trend from the beginning of the measurement period until about 2006, followed by a rapid recovery and continued high inter-annual variability through 2016 (Figure 2.12b). The trend in $\delta^{13}C$-$CH_4$ prior to 2000 indicated that the increase in atmpospheric concentrations was due to thermogenic (fossil) $CH_4$ emissions; the reversal of this trend after 2007 indicates a shift to biogenic sources (Figure 2.12c). However, the concurrent increase in atmospheric ethane after 2007 is a contradiction as this indicates a thermogenic source, particularly in the northern hemisphere (Schwietzke et al. 2014; Hausmann et al. 2015).

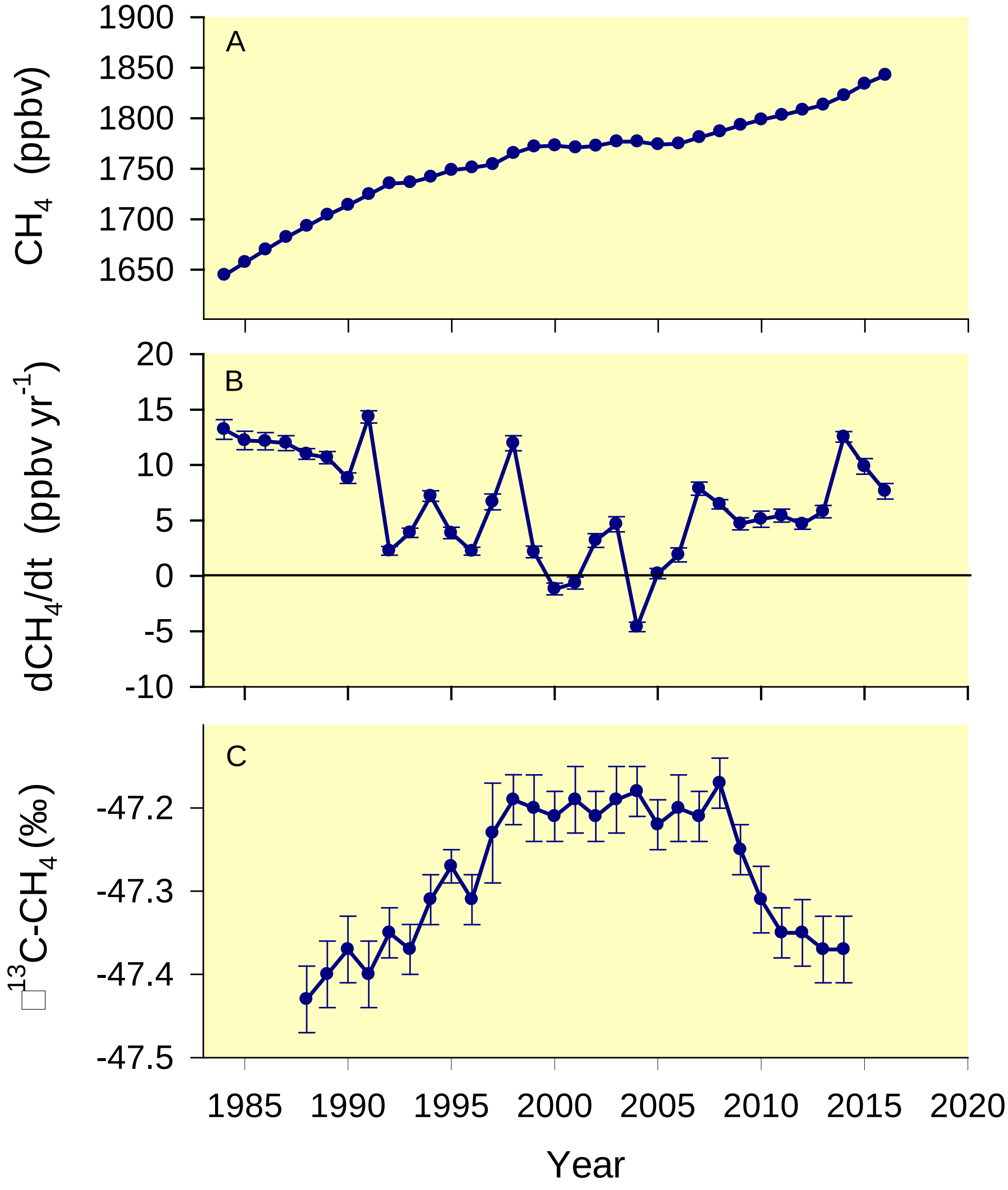


**Figure 2.12 Globally averaged atmospheric $CH_4$ mixing ratios (Frame A) and instantaneous rates of change (Frame B) and C isotope signature variation (Frame C) Data sources: NOAA/ESRL (www.esrl.noaa.gov/gmd/ccgg/trends_ch4/)(Dlugokencky et al. 1994) and Schaefer et al. (2016)**

Understanding the underlying causes of temporal variation in atmospheric $CH_4$ concentrations is an active area of research. To estimate temporal emission trends, Bergamaschi et al. (2013) used column averaged $CH_4$ mixing ratios from the Scanning Imaging Absorption Spectrometer for Atmospheric Cartography (SCIAMACHY) on board Envirosat for atmospheric inversion sensitivity experiments. The modelled global emissions showed small anomalies between 2000 and 2006 (±10 Tg $yr^{-1}$). There was a significant increase after 2006 and emissions between 2007 and 2010 were between 16 and 20 Tg $yr^{-1}$. The increase was mainly attributed to anomalies in the tropics (9-14 Tg $y^{-1}$) and the mid-latitude Northern Hemisphere (6-8 Tg $yr^{-1}$). Almost 30% of the increase in anthropogenic emissions in the Emissions Database for Global Atmospheric Research (EDGAR) v4.3.2 dataset was attributed to China – 6Tg from coal mining and 1.7 Tg from agriculture (primarily rice cultivation and enteric fermentation). The inversion estimate by Bergamaschi et al. (2013) attributed about one third less emissions to China. Superimposed on the rising emissions trend were significant inter-annual variations attributed to wetlands (±10 Tg $yr^{-1}$) and biomass burning (±7 Tg $yr^{-1}$).

Several studies suggested that inter-annual variability of $CH_4$ growth was driven mostly by variations in natural emissions from wetlands (Rice et al. 2016; Bousquet et al. 2006; Bousquet et al. 2011). These modelling efforts suggested that tropical wetlands were responsible for between 50 and 100% of the inter-annual fluctuations, but results between the models were inconsistent for the magnitude and geographic distribution of the wetland source. Bousquet et al. (2011b) suggested that increased emissions from wetlands accounted for the reprise of growth in atmospheric concentrations and they concluded that OH variation over the period analysed accounted for <1% of the variation. Pison et al. (2013) used two atmospheric inversion models and the ORCHIDEE model and found greater uncertainty in the role of wetlands in inter-annual variability between 1990 and 2009 and during the 1999-2006 pause. Poulter et al. (2017) used several biogeochemical models and inventory-based wetland area data to show that wetland $CH_4$ emissions increases in the boreal zone have been offset by decreases in the tropics and concluded that wetlands have not contributed significantly to renewed atmospheric $CH_4$ growth.

The development of credible time series of methyl chloroform ($CH_3CCl_3$) observations offered a way understand temporal dynamics of OH abundance and applying this to global budgets further weakened the argument for the role of wetlands in determining temporal trends since 1990 (Rigby et al. 2013; McNorton et al. 2016). McNorton et al. (2016) showed that changes in the atmospheric OH sink explained a large portion of the suppression in global $CH_4$ concentrations relative to the pre-1999 trend. Turner et al. (2017) also looked at this variability and found that there was a 35 Tg $yr^{-1}$ increase in $CH_4$ emissions between 1993 and 2003, the majority of which was found to be in the Northern Hemisphere. This was accompanied by a 7% increase in global mean OH between 1991 and 2000. They attribute the 1999-2006 stabilisation to slowing of the increase of $CH_4$ emissions after 1998 and the enhanced OH sink. The reprise of growth was the result of a decrease in $CH_4$ emissions accompanied by a decrease in the OH sink.

Changes in the isotopic signature of $CH_4$ in the atmosphere suggests a shift from fossil-fuel to biogenic sources (Schaefer et al. 2016; Schwietzke et al. 2016) (*moderate evidence, high agreement*). The depletion of $\delta^{13}C_{atm}$ beginning in 2009 could be due to changes in several sources. Lower fire emissions combined with higher tropical wetland emissions could explain the $\delta^{13}C$ perturbations to atmospheric $CH_4$ sources (Schaefer et al. 2016; Worden et al. 2017). However, because tropical wetland emissions are higher in the Southern Hemisphere, and the remote sensing observations show that $CH_4$ emissions increases are largely in the north tropics (Bergamaschi et al. 2013; Melton et al. 2013; Houweling et al. 2014), an increased wetland source does not fit well with the southern hemisphere $\delta^{13}C$ observations. Schaefer et al. (2016) suggested that agriculture is a more probable source of increased emissions, and particularly livestock in the tropics, which is consistent with inventory data. Patra et al. (2016) suggested that the renewed growth was associated with increases in livestock herds and other agricultural sources.

The importance of fugitive emissions in the global atmospheric accumulation rates appears to be growing (*medium evidence, high agreement*). The increased production of natural gas in the US from the mid 2000's is of particular interest because it coincides with the reprise of atmospheric $CH_4$ growth. Rice et al. (2016) used measurements of the stable isotopic composition of atmospheric $CH_4$ ($^{13}C/^{12}C$ and D/H) in the Northern Hemisphere between 1977 and 2009 to constrain the sources of $CH_4$ and found an increase in fugitive fossil fuel emissions since 1984 with most of this growth occurring after 2000. Hausmann et al. (2015) found a significant ethane–methane correlation between 2007 and 2014, which was absent between 1999 and 2006 and estimated that fugitive emissions during the reprise of growth increased by between 25 and 45 Tg $yr^{-1}$. Reconciling the increased fugitive emissions with increased isotopic depletion of atmospheric $CH_4$ indicates that there were likely multiple changes in emissions and sinks that affected atmospheric accumulation.

With respect to atmospheric $CH_4$ growth rates, we conclude that there is significant and ongoing accumulation of $CH_4$ in the atmosphere (*very high confidence*). The reason for the pause in growth rates and the subsequent reprise appear to be at least partially associated with land use and land use change. Contrary to the findings of AR5, wetlands are not the primary drivers of inter-annual variability or the cause of the pause in growth rates in the early 2000s (*medium evidence, high agreement*). We also conclude that variation in the atmospheric OH sink plays an important role in the year to year variation of the $CH_4$ growth rate, but does not explain the entirety of the changes in the growth rates (*medium evidence, medium agreement*). Fugitive emissions *likely* contribute to the reprise in growth after 2006 (*medium evidence, high* *agreement*) Additionally, the recent depletion of $^{13}C$ isotope in the atmosphere indicates that growth in biogenic $CH_4$ sources explains part of the current growth (*high agreement, robust evidence*); however, attributing these sources to different activities like livestock and the expansion of rice cultivation is still difficult. There is some evidence that increases in other tropical sources may be playing a role in the reprise of the growth rate (*low agreement, limited evidence*).

### *2.4.2.3 Global CH4 budget*

AR5 presented decadal global $CH_4$ budgets beginning in 1980; the Kirschke et al. (2013) budget in Table 2.1 represents the most recent decade reported. A new budget has been developed, covering the period 2000 to 2012 (Saunois et al. 2016). We present the revised budget for the final decade reported in AR5 and for the final year of the new analysis. The main sources of $CH_4$ are natural emissions from wetlands and anthropogenic sources, with significant emissions from agriculture, forestry and other land use. Each of these sources is greater than the net land sink in $CO_2$ equivalents, using the AR5 global warming potential value of 28 (Bastviken et al. 2011). Global emissions are between 600 and 700 Tg $CH_4$ $yr^{-1}$ and 60-70% of this is due to anthropogenic sources (Kirschke et al. 2013; Bruhwiler et al. 2014; Janssens-Maenhout et al. 2017). The primary sink for atmospheric $CH_4$ is consumption by tropospheric OH; stratospheric reactions with chlorine and atomic oxygen radicals, and consumption in soils by methanotrophic bacteria are minor sinks. However, these minor sinks are 3 to 5 times greater than the current rate of annual increase of $CH_4$ in the atmosphere, so changes to them could affect atmospheric accumulation rates. Increasing atmospheric concentrations are largely driven by anthropogenic emissions, which appear to be increasing. Estimates derived from inverse modelling vary, and they suggest that the current annual rates of increase are between 6 and 14 Tg $CH_4$ $yr^{-1}$ between 2000 and 2012 (Kirschke et al. 2013; Saunois et al. 2016). Observations show that annual rates of increase have varied between 3.84 and 10.30 Tg since 2010 (NOAA/ESRL, www.esrl.noaa.gov/gmd/ccgg/trends_ch4*)*.

**Table 2.2 Bottom-up and top-down estimates of the components of the global $CH_4$ budget by source type (Tg $CH_4$ $yr^{-1}$) for 2000-2009 and 2012. The numbers in brackets represent the minimum and maximum values in reported studies. The atmospheric annual increase reported is the assumed value for inversions that do not report the global sink**

| | **Kirschke et al. 2013** | | **Saunois et al. 2016** | | | |
|---|---|---|---|---|---|---|
| | 2000-2009 | | 2000-2009 | | 2012 | |
| | Bottom Up | Top-down | Bottom-up | Top-down | Bottom-up | Top-down |
| **Natural sources** | **347 (238–484)** | **218 (179–273)** | **382 (255–519)** | **234 (194–292)** | **386 (259–532)** | **221 (192–302)** |
| Natural wetlands | 217 (177–284) | 175 (142–208) | 183 (151–222) | 166 (125–204) | 187 (155–235) | 172 (155–201) |
| Fresh water | 40 (8–73) | | 122 (60–80) | | | |
| Wildlife | 15 (15–15) | | 10 (5–15) | | | |
| Termites | 11 (2–22) | | 9 (3–15) | | | |
| Wildfires | 3 (1–5) | | 3 (1–5) | | | |
| Permafrost | 1 (0–1) | | 1 (0–1) | | | |
| Non-land based) | 61 (35 – 85) | | 68 (40–106) | | | |
| **Anthropogenic sources** | **331 [304–368]** | **335 [273–409]** | **338 [329–342]** | **319 [255–357]** | **370 [351–385]** | **347 [262–384]** |
| Agriculture and waste | | 209 [180–241] | | 183 [112–241] | | 200 [122–213] |
| Enteric fermentation & manure management | 101 [98–105] | | 103 [95–109] | | 107 [100–112] | |
| Landfills & waste | 63[56–79] | | 57 [51–61] | | 60 [54–66] | |
| Rice cultivation | 36 [33–40] | | 29 [23–35] | | 29 [25–39] | |
| Biomass burning | 35 [32–39]* | 30 [24–45]* | 18 [15–20] | | 17 [13–21] | |
| Non-land based | 96 [85–105] | 96 [77–123] | 112 [107–126]* | 136 [93–179]* | 134 [123–141]* | 147 [118–188]* |
| **Sinks** | | | | | | |
| Soils | 28 [9–47] | 32 [26–42] | | 32 [27–38] | | 36 [30–42] |
| Atmospheric chemical loss | 604 [483–738] | 518 [510–538] | | 514 | | 518 |
| **TOTALS** | | | | | | |
| Sum of sources | 678 [542–852] | 548 [526–569] | 719 [583–861] | 552 [535–566] | 756 [609–916] | 568 [542–582] |
| Sum of sinks | 632 [592–785] | 540 [514–560] | | 546 | | 555 |
| Imbalance (sources−sinks) | | 8 [−4–19] | | 6 | | 14 |
| Atmospheric growth rate | | 6 | | 6.0 [4.9-6.6] | | 14.0 [] |

* Includes biofuel burning

### *2.4.2.4 Land use effects*

There are several datasets that are typically used for tracking emissions for agriculture, forestry and other land use (AFOLU). In Figure 2.13, we present national greenhouse gas inventory data, EDGAR (Emissions Database for Global Atmospheric Research) and FAOSTAT (Food and Agriculture Organization Corporate Statistical Database). Whereas there is generally good agreement between these datasets for agriculture (Roman-Cuesta et al. 2016), we can conclude that in the agricultural sector, emissions are higher in non-Annex 1 countries than in Annex 1 countries (*high confidence*). Increasing growth of agricultural emissions in the tropics is due to many factors including population growth and increases in trade and export of agricultural commodities in many countries (le Polain de Waroux et al. 2017; Meyfroidt et al. 2014; Geist and Lambin 2002) .

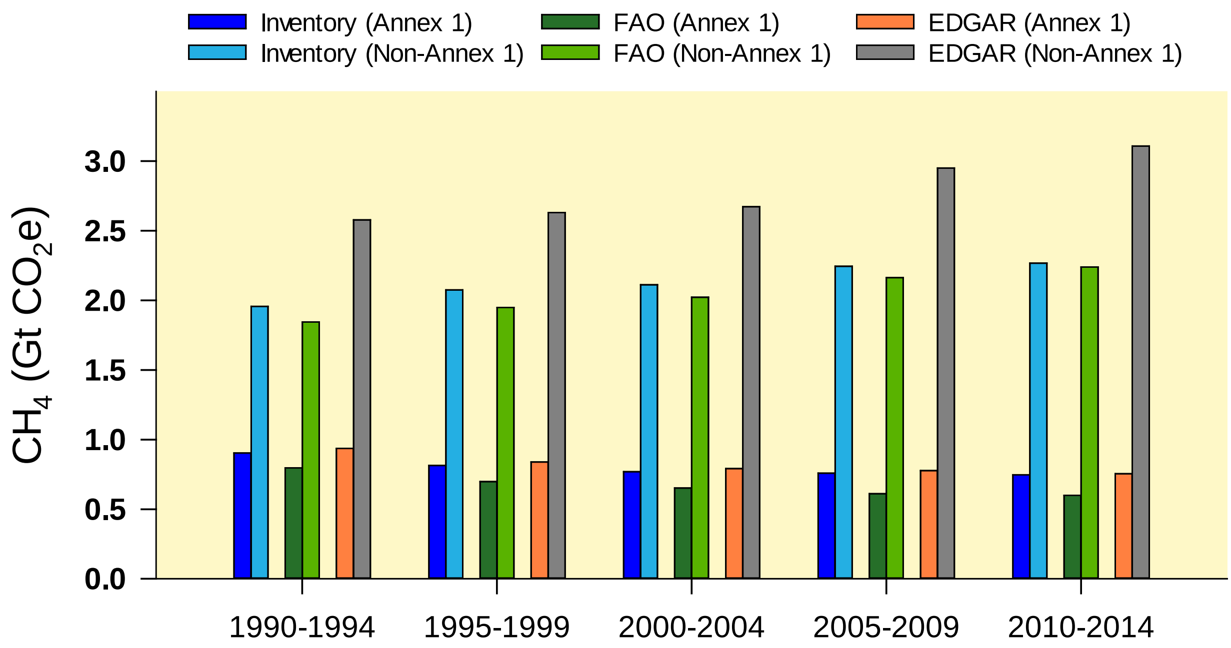


**Figure 2.13 Agricultural $CH_4$ emissions for Annex 1 and Non-Annex 1 countries from national GHG inventory data, FAOSTAT (FAO 2015), and EDGAR databases (Janssens-Maenhout et al. 2017b)**

Agricultural emissions are predominantly from enteric fermentation and rice, with manure management and waste burning contributing small amounts (Figure 2.14). Livestock production is responsible for 33% of total global emissions and 66% of agricultural emissions (Source: EDGAR 4.3.2 database, accessed May 2018). Most of the livestock emissions are from developing countries (EDGAR 4.3.2, (EPA 2013; Tubiello et al. 2014)). Asia has the largest livestock emissions (37%) and emissions in the region have been growing by around 2% per year. Africa is responsible for only 14%, but emissions are growing fastest in this region at around 2.5%. In Latin America and the Caribbean, livestock emissions are decreasing at around 1.6% per year and the region makes up 16% of emissions. Developed countries are responsible for about 17% of emissions and these are decreasing by about 1.5% per year. Rice emissions are responsible for about 24% of agricultural emissions, and 89% of these are from Asia. Rice emissions are increasing by 0.9% per year in that region. These trends are predicted to continue through 2030 (EPA 2013).

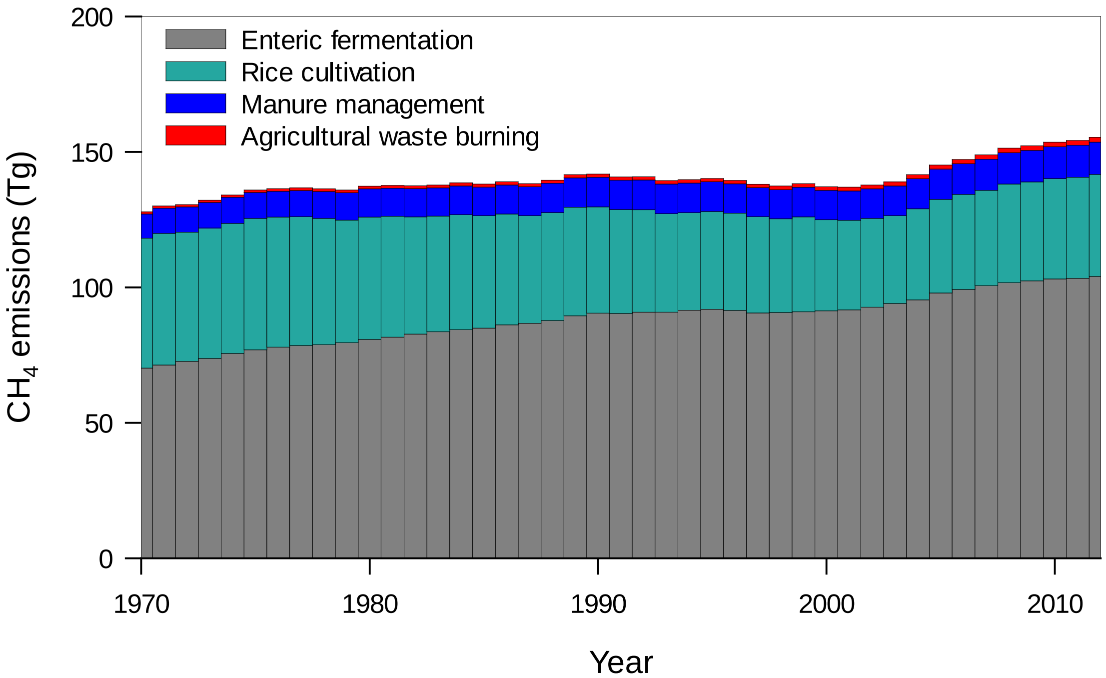


**Figure 2.14 Agricultural $CH_4$ emissions between 1970 and 2012. Source: Edgar database 4.3.2**

Upland soils are a net sink of atmospheric $CH_4$, but soils both produce and consume the gas. The net soil–atmosphere flux is the result of the balance between the two offsetting processes of methanogenesis (microbial production) and methanotrophy (microbial consumption) (Serrano-Silva et al. 2014). Microbial consumption requires aerobic conditions because the biochemical process requires a monooxygenase enzyme. Methanogenesis is the process of microbial production of $CH_4$ in anaerobic conditions. Methanogenesis is an important process in wetland soils and rice paddies and these systems are usually sources of $CH_4$ for the atmosphere. However, methanogenesis can also occur in upland soils in anaerobic 'microsites' inside soil aggregates. Methanotrophy is the dominant process in upland soils, where oxidation generally exceeds production. Methanotrophy is also an important process in wetland and rice paddy soils at the oxic soil-water interface and in the rhizosphere, and this limits the amount of $CH_4$ emitted by these soils. Between 40% and 80% of the $CH_4$ that diffuses through the oxic zones in soils and sediments is consumed therein (Laanbroek, 2018; Serrano-Silva et al. 2014).

On the global scale climatic zone, soil texture, and land cover have an important effect on $CH_4$ 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 $CH_4$ than medium and fine textured soils, and forests take up more than other ecosystems. Low levels of nitrogen fertilisation can stimulate soil $CH_4$ uptake, while higher fertilisation rates decrease uptake (Edwards et al. 2018). The effect of N additions is cumulative and repeated fertilisation events have progressively greater suppression effects. Zhuang et al. (2013) estimated that between 1998 and 2004, that N fertilisation suppressed $CH_4$ oxidation by 26 Tg. Soil $CH_4$ consumption has been increasing during the second half of the 20$^{th}$ century and it is expected to continue to increase by as much as 1 Tg in the 21$^{st}$ century.

Northern peatlands (40°-70°N) constitute a significant source of atmospheric $CH_4$, emitting about 48 Tg $CH_4$, or about 10% of the total emissions to the atmosphere (Zhuang et al. 2006; Wuebbles and Hayhoe 2002). $CH_4$ emissions from natural northern peatlands are highly variable with the highest rate from fen ecosystems. The rate of $CH_4$ emissions from natural peatlands depends on many factors including water table depth, temperature, vegetation (direct release via vascular plants as well) and other factors. Under the climate change, interactions of these complex factors will be the main determinant of emissions from northern peatlands. However, management of undisturbed peatlands, as well as the restoration of disturbed ones, alter the exchange of $CH_4$ with the atmosphere. Abdalla et al. (2016) reviewed 87 studies with paired observations on drained and undrained sites and found that on average drainage reduced $CH_4$ emissions by 84%. They also reviewed 16 sites with restoration activities and found that rewetting increased emissions by 46% above pre-drainage levels. The study did not resolve how $CH_4$ fluxes change over time with drainage or rewetting; however, Vanselow-Algan et al. (2015) showed that high $CH_4$ emissions in rewetted bogs can persist for at least 30 years. Most direct uses of northern peatlands, such as peat extraction, agriculture and forestry require drainage. Lowering the water table usually turns peat soils from $CH_4$ sources to sinks as a result of reduced methanogenesis in the waterlogged peat and enhanced methanotrophy in the aerated zone of the surface peat (Joosten and Couwenberg 2008; Nayyar and Jindal 2010). Drained peatlands which are usually considered as negligible or "zero" methane sources, still emit $CH_4$ under wet weather conditions and especially in drainage ditches (Drösler et al. 2013; Sirin et al. 2012), which cover only a small percentage of the drained area. In some cases drainage ditch emissions are so high that drained peatlands are comparable to natural ones (Sirin et al. 2012; Wilson et al. 2016).

Because of the large uncertainty in the tropical peatland area, estimates of the global flux are highly uncertain. Hergoualc'h and Verchot (2012) conducted a meta-analysis on peat $CH_4$ fluxes before and after land use change. Conversion of primary forest to rice production increased emissions from 29±10 kg $CH_4$-C $ha^{-1}$ $y^{-1}$ to 108±60 kg $CH_4$-C $ha^{-1}$ $yr^{-1}$. For land uses that required drainage, emissions decreased to 9.5+6.1 kg $CH_4$-C $ha^{-1}$ $yr^{-1}$. Methane fluxes displayed an exponential response to water table depth changes across all land uses. There are no representative measurements of emissions from drainage ditches in tropical peatlands.

#### *2.4.2.5 Future trends and climate change effects*

Climate change is expected to impact the key terrestrial biogeochemical cycles, via an array of complex feedback mechanisms that will act to either enhance or decrease future, $CH_4$ and $N_2O$ emissions (described in Section 2.2 and 2.3). The balance of these positive and negative feedbacks remains uncertain. Estimations from climate models included in AR5, CMIP5 and $C^4$MIP exhibit large differences for the different carbon and nitrogen cycle feedbacks and how they change in a warming climate (Anav et al. 2013; Friedlingstein et al. 2006; Friedlingstein et al. 2014). The differences are in large part due to the uncertainty regarding how primary productivity will evolve, with many of the models not even agreeing on the sign of change. Furthermore, many models do not include a nitrogen cycle, which limits the $CO_2$ fertilisation effect. These uncertainties are further exacerbated by the lack of observational constraints (Prentice et al. 2015a).

The future of the $CH_4$ budget will depend in large part on changes in the atmospheric OH sink. Methane trajectories in AR5 were based on calculations of the MAGICC model that accounted for changes in the OH sink, anthropogenic CO emissions, $NO_x$, VOCs, temperature, and the negative feedback of increasing $CH_4$ concentrations on OH (Meinshausen et al. 2011). A more recent analysis based on three chemical transport models showed that atmospheric lifetime of $CH_4$ is likely to increase through about 2070 and decrease slightly thereafter (Holmes et al. 2013). While the trajectory is similar to that in AR5, the atmospheric lifetime is predicted to be approximately one year shorter in the new analysis. Sensitivity analyses of the models lead to an upward revision in the 100-yr GWP estimate to 32. This study assumes that all other sinks remain constant, but other analyses suggest that increased atmospheric abundance will also increase the global soil sink. Similar results were found in simulations performed for the Atmospheric Chemistry and Climate Modeling Intercomparison Project (ACCMIP) (Voulgarakis et al. 2013). Mean tropospheric lifetime for the year 2000 was estimated to be 9.8±1.6 $yr^{-1}$, which was lower than AR5 estimates. Future projections were made using the four Representative Concentration Pathways (RCPs). Decreases in global methane lifetime of 4.5±9.1% were projected for RCP 2.6, the scenario with lowest radiative forcing by 2100. Simulation with the high radiative forcing RCP8.5 produced increases of 8.5±10.4%. In this latter scenario, atmospheric $CH_4$ concentration was the key driver of the evolution of OH and methane lifetime because $CH_4$ concentration more than doubled by 2100, which resulted in significantly greater consumption of atmospheric OH.

There are few studies on the effects of climate change on the land sources and sinks of $CH_4$ from which to assess the future trends. From a mechanistic standpoint, we have known for a long time that gas diffusion is the major rate-limiting step of soil methane uptake (Striegl 1993) (*robust evidence, high agreement*). In areas of the world where climate change leads to higher temperatures and drier soils in upland ecosystems, climate change will enhance the soil $CH_4$ sink (Curry 2009; Rowlings et al. 2012; Luo et al. 2013).

Increased atmospheric $CH_4$ will also increase the diffusion gradient leading to increased soil uptake by as much as 15% (Curry 2009).

Soil sources in wetlands and rice paddies could also be affected by climate change. The extent to which permafrost melts and creates new wetlands in high latitudes, and wetlands in other regions expand or contract as a result of changes in rainfall and temperature, these sources could increase in some regions and decrease in others (Ringeval et al. 2010; Schuur et al. 2015). A warming climate and conversion of the land surface can influence permafrost and wetland ecosystems. Thawing permafrost and degradation of wetlands is expected to increase atmospheric greenhouse gases and act as a positive feedback. Although current wetland methane emissions remain uncertain, they may account for up to 40% of total global methane emissions (Saunois et al., 2016), and for much of annual variability (McNorton et al., 2016).

Climate change may affect enteric fermentation emissions through impacts on forage quality. Increased $CO_2$ results in decreased N content and increases nonstructural carbohydrates in some forages. Reduced water availability can increase digestibility in grasses by increasing N concentrations, but there is a productivity tradeoff. Cool-season forages with a $C_3$ photosynthetic pathway will benefit from elevated $CO_2$, either because of increased photosynthesis or decreased soil moisture depletion due to stomatal closure. Elevated $CO_2$ can also stimulate N fixation (Dumont et al. 2015; Thivierge et al. 2016).

## 2.4.3 Nitrous Oxide

### *2.4.3.1 Atmospheric trends*

The atmospheric abundance of $N_2O$ has increased since 1750, from a pre-industrial concentration of 270 ppbv to 330 ppbv in 2017 (U.S. National Oceanagraphic and Atmsopheric Agency, Earth Systems Research Laboratory; Figure 2.15). The rate of increase has also increased, from approximately 0.15 ppbv year$^{-1}$ 100 years ago, to 0.85 ppbv year$^{-1}$ over 2001-2015 (Wells et al. 2018). Recent measurements of isotopic $N_2O$ composition ($^{14/15}$N) show a decrease in the $\delta^{15}$N to $N_2O$ ratio between 1940 and 2005, which confirms that consumption of synthetic nitrogen (N) fertilizer is largely responsible for the observed increase in $N_2O$ concentrations (Park et al. 2012). Increased nitrogen deposition and climate warming have also contributed, particularly since 1980(Tian et al. 2016). The increase in atmospheric $N_2O$ concentrations is concerning not only because $N_2O$ is responsible for approximately 6% of global radiative forcing from anthropogenic greenhouse gases; its ozone-depletion potential-weighted emissions of 0.47 Mt CFC-11-equivalent in 2008 outweighs the sum of emissions from all other ozone-depleting substances controlled by the Montreal Protocol (Saikawa et al., 2014; WMO, 2014). Moreover, as noted in AR5, the long atmospheric lifetime of $N_2O$ (118-131 years) means that atmospheric concentrations would take more than a century to stabilize following the stabilization of global emissions (Ciais et al. 2013).

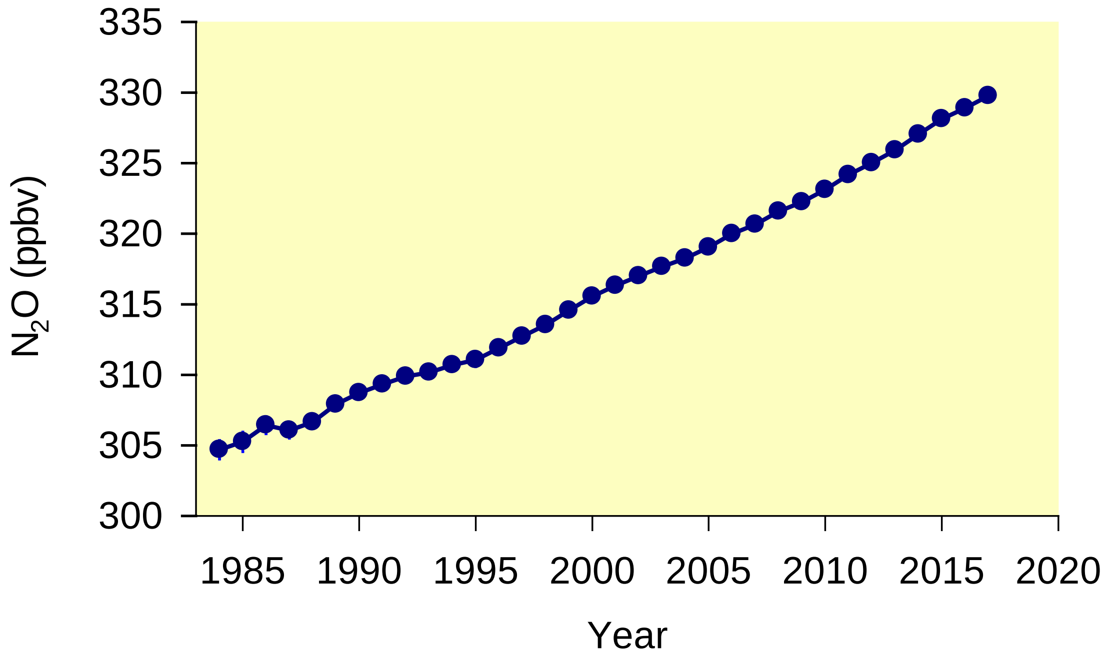


**Figure 2.15 Globally averaged atmospheric $N_2O$ mixing ratios using combined data from the NOAA/ESRL Global Monitoring Division (https://www.esrl.noaa.gov/gmd/hats/combined/N2O.html)**

#### *2.4.3.2 Global $N_2O$ budget*

Recent estimates using inversion modelling and process models estimate total global $N_2O$ emissions of 15.3 to 17.3 (bottom-up) and 15.9 to 17.7 Tg N (top-down), demonstrating relatively close agreement (Davidson and Kanter, 2014; Tian et al. 2018; Wells et al., 2018). Microbial denitrification and nitrification processes are responsible for more than 80% of total global $N_2O$ emissions, which includes natural soils, agriculture, and oceans, with the remainder coming from non-biological sources such as biomass burning and fossil-fuel combustion (Fowler et al. 2015). A recent development since AR5 is the ability to combine these methodologies using a multi-inversion approach and an ensemble of surface observations to better constrain the regional and temporal distribution of emissions (Saikawa et al. 2014; Wells et al. 2018).

$N_2O$ has both natural and anthropogenic sources (Table 2.2). Natural emissions have terrestrial, marine and atmospheric sources. Recent estimates of terrestrial sources suggest a higher and slightly more constrained emissions range than reported in AR5: approximately 9 (7-11) Tg $N_2O$-N year$^{-1}$ (Saikawa et al. 2014; Tian et al. 2016) versus 6.6 (3.3-9.0) Tg $N_2O$-N year$^{-1}$ (Ciais et al. 2013). Similarly, recent estimates of marine $N_2O$ emissions (2.5 ± 0.8 Tg $N_2O$-N year$^{-1}$; Buitenhuis et al., 2017; 4.6 ± 0.3 Tg $N_2O$-N year$^{-1}$; Saikawa et al., 2014) show a more well constrained range than AR5, although the AR5 estimate (3.8 Tg $N_2O$-N year$^{-1}$ with uncertainty bounds of 1.8-9.4 Tg $N_2O$-N year$^{-1}$) is within the range.

**Table 2.3 Annual $N_2O$ inventories by sector, all units in Tg (Source: Davidson and Kanter, 2014)**

| | FAO | EDGAR | EPA 2012 |
|---|---|---|---|
| Agriculture | 4.1 | 3.8 | 4.6 |
| Fertilizer | 1.4 | | |
| Direct | 1.1 | | |
| Indirect | 0.3 | 3.6 | |
| Manure | 1.8 | | |
| Direct | 1.4 | | 4.2 |
| Indirect | 0.4 | | |
| Organic soils | 0.2 | | |
| Crop residues | 0.3 | | |
| Manure management | 0.3 | 0.2 | 0.4 |
| Biomass burning | | 1.1 | |
| Residue burning | 0.01 | | 1.6 |
| Other | | | 0.1 |
| Industry, energy and transport | | 1.7 | 0.9 |
| Wastewater | | 0.2 | 0.2 |
| Solvent and other product use | | | 0.2 |
| Total | 11.31 | 14.8 | 10.8 |

Both top-down and bottom-up approaches can differentiate natural from anthropogenic N2O contributions (Davidson and Kanter 2014). For the top-down analyses, changes in the atmospheric abundance of N2O from pre-industrial to the present are assumed to be entirely anthropogenic. Natural emissions are assumed to have remained stable over this period (~11 Tg N $yr^{-1}$) and are subtracted from the total to yield an estimate of anthropogenic emissions. The bottom-up approach uses protocols developed by the IPCC that, in their simplest and most widely applied form, multiply measures of activity in agriculture, energy generation, industry and other sectors by emission factors (EFs) to estimate the $N_2O$ emitted per unit of activity (de Klein et al. 2014). Recent estimates using both approaches suggest net anthropogenic emissions of approximately 5.3 Tg $N_2O$-N $year^{-1}$ (Davidson and Kanter 2014). This estimate also accounts for lower tropical forest soil emissions of approximately 0.9 Tg $N_2O$-N $year^{-1}$ as a result of deforestation, both past and present (Davidson 2009).

### *2.4.3.3 Land use effects*

Agriculture is the predominant source of anthropogenic $N_2O$ and is responsible for approximately two-thirds of emissions (Janssens-Maenhout et al. 2017a). Recent studies estimate emissions of 4.1 Tg $N_2O$-N $year^{-1}$ (3.8-6.8 Tg $N_2O$-N $year^{-1}$; Oenema et al., 2014). Total emissions from this sector are the sum of direct and indirect emissions. Direct emissions from soils are the result of mineral fertilizer and manure application, manure management, deposition of crop residues, cultivation of organic soils and biological nitrogen fixation. Indirect emissions come from downstream and downwind water bodies and soils after nitrate has been leached or nitrogen oxides and ammonia emissions have been deposited back on agricultural land. The main driver of agricultural $N_2O$ emissions (and other agricultural N losses) is a lack of synchronization between crop N demand and soil N supply, with approximately 50% of N applied to agricultural land not taken up by the crop (Zhang et al. 2017a). Recent findings at regional scales confirm significant increases in agricultural $N_2O$ emissions from the Asian agricultural sector in recent years, driven by an increase in synthetic N fertilizer use (Saikawa et al. 2014; Wells et al. 2018)

Agricultural $N_2O$ emissions (and soil $N_2O$ emissions generally; Figure 2.16) are characterized by hot spots and hot moments (Groffman et al. 2009) meaning that they are often concentrated in brief periods and small areas where conditions are optimal (e.g. high soil moisture after springtime N application). Since AR5, our understanding of these conditions has improved, particularly with regards to soil rewetting and freeze-thaw cycles outside of the growing season (Liu et al. 2018b; Wagner-Riddle et al. 2017). Between 35% and 65% of total annual $N_2O$ emissions from terrestrial sources may result from thaw-related fluxes, as a result of increased substrate availability, changes in denitrifying enzymes, and the release of previously produced $N_2O$. Neglecting these emissions could lead to an underestimation of global agricultural $N_2O$ emissions by 17%-28% (Wagner-Riddle et al. 2017).

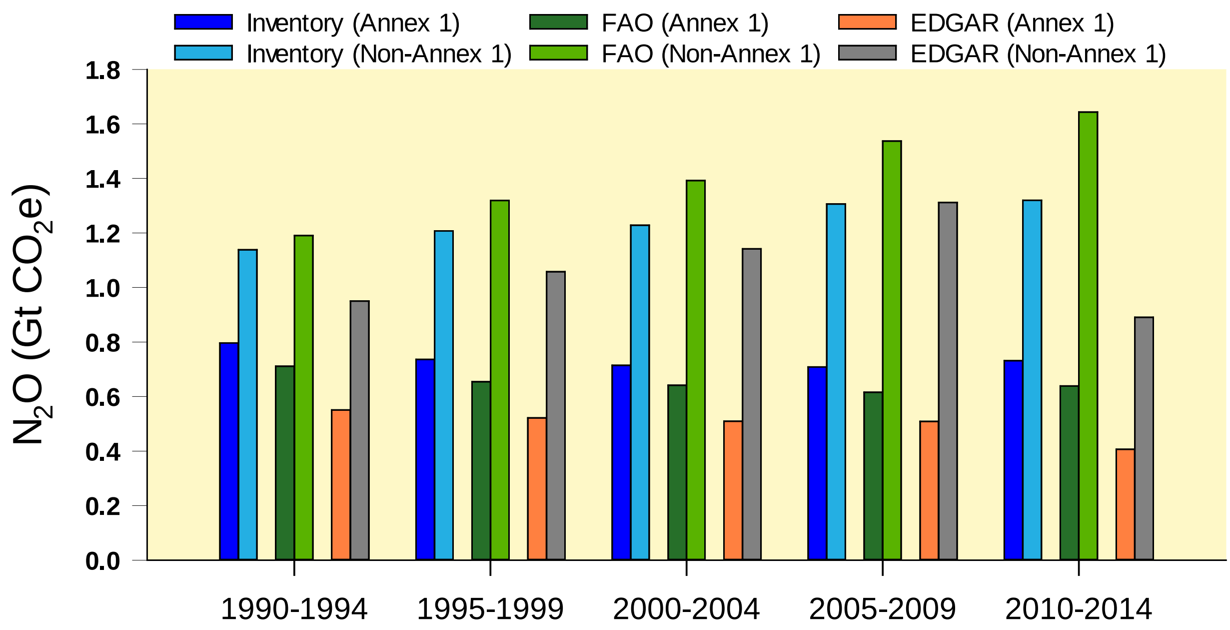


**Figure 2.16 Agricultural $N_2O$ emissions for Annex 1 and Non-Annex 1 countries from national GHG inventory data, FAOSTAT, and EDGAR databases**

Industry and fossil fuel combustion is the largest non-agricultural source of anthropogenic $N_2O$ emissions, responsible for approximately 0.9 Tg $N_2O$-N year$^{-1}$ (0.7-1.6 Tg $N_2O$-N year$^{-1}$) or 15% of total gross anthropogenic $N_2O$ emissions (Wiesen et al. 2013). Nitric and adipic acid production are the major industrial sources, while stationary combustion (mainly from coal power plants) is the energy sector's main source. In both cases $N_2O$ emissions are the result of the oxidation of atmospheric $N_2$ and organic N in fossil fuels. Biomass burning is responsible for approximately 0.7 Tg $N_2O$-N year$^{-1}$ (0.5-1.7 Tg $N_2O$-N year$^{-1}$) or 11% of total gross anthropogenic emissions due to the release of $N_2O$ from the oxidation of organic N 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. Emissions from wastewater are approximately 0.2 Tg $N_2O$-N year$^{-1}$ or 3% of total gross anthropogenic emissions, emitted either directly from wastewater or wastewater management facilities (Bouwman et al. 2013). Aquaculture, while currently responsible for less than 0.1 Tg $N_2O$-N year$^{-1}$, is one of the fastest growing sources of anthropogenic $N_2O$ emissions (Williams and Crutzen, 2010; Bouwman et al. 2013). Finally, increased N deposition onto the ocean is estimated to have increased the oceanic $N_2O$ source by 0.2 Tg $N_2O$-N year$^{-1}$ or 3% of total gross anthropogenic emissions (Suntharalingam et al. 2012).

We conclude that $N_2O$ is continuing to accumulate in the atmosphere at an increasingly higher rate (*very high confidence*), driven primarily by increases in manure production and synthetic N fertilizer use from the mid-20$^{th}$ century onwards (*high confidence*). Findings since AR5 have constrained regional and global estimates of annual $N_2O$ emissions and improved our understanding of the spatio-temporal dynamics of $N_2O$ emissions, with soil rewetting and freeze-thaw cycles important determinants of total annual emission fluxes (*medium confidence*).

Studies since AR5 highlight two major uncertainties in the estimation of anthropogenic $N_2O$ emissions using bottom-up methods, particularly from the agricultural sector: emission factors (EFs) and indirect emissions. First, the Tier 1 EFs assume a linear relationship between N application rates and $N_2O$ emissions, with a 1% EF applied to synthetic N fertilizer rates to estimate direct emissions. However, recent studies are increasingly finding nonlinear relationships, suggesting that $N_2O$ emissions per hectare are lower than the

Tier 1 EFs at low N application rates, and higher at high N application rates – a result of greater excess N unused by crops, which is then available to be emitted as $N_2O$ (Shcherbak et al. 2014; Van Lent et al. 2015; Satria 2017). For example, applying the IPCC Tier 1 EF to a 50 kg N $ha^{-1}$ reduction in N application rate would generate an estimated reduction in $N_2O$ emissions of 0.5 kg $N_2O$-N $ha^{-1}$, regardless of the initial application rate. However, using a nonlinear EF for upland grain crops derived via meta-analysis, a reduction from 50 kg N $ha^{-1}$ to zero would reduce emissions by 0.37 kg $N_2O$-N $ha^{-1}$, while a reduction from 300 kg N $ha^{-1}$ to 250 kg N $ha^{-1}$ would reduce emissions by 0.84 kg $N_2O$-N $ha^{-1}$, suggesting greater mitigation potential in regions with higher N application rates. This not only has implications for how agricultural $N_2O$ emissions are estimated in national and regional inventories, it also suggests that in regions of the world where low N application rates dominate, such as sub-Saharan Africa and parts of Eastern Europe, relatively large increases in N fertilizer use would generate relatively small increases in agricultural $N_2O$ emissions. Other factors that impact EF magnitude include crop and fertilizer type, soil carbon, soil water content, pH, mean annual temperatures, and organic amendment type – though significant uncertainties remain, particularly for tropical soils (Cayuela et al. 2017; Charles et al., 2017; Meurer et al. 2016; Shcherbak et al., 2014). Nevertheless, there is evidence that errors in emission estimates from applying the Tier 1 EF at small scales are largely cancelled when aggregated to larger scales (Del Grosso et al. 2010).

The second major uncertainty in estimating agricultural $N_2O$ emissions comes from indirect emissions. Recent studies suggest that the Tier 1 EFs are low, especially the 0.75% EF for indirect $N_2O$ from leached nitrate. One study in the U.S. Corn Belt estimates an EF closer to 2% and emissions are highly dependent on stream hierarchy, which would imply an underestimation of current indirect emissions of up to nine fold and translate to a total underestimation of agricultural $N_2O$ emissions in the region of up to 40% (Turner et al. 2015). The gap between estimated and actual EFs will become increasingly important in a changing climate, as described below.

To conclude, findings since AR5 increasingly highlight the limits of linear $N_2O$ emission factors, particularly from field to regional scales, with emissions rising nonlinearly at high N application rates (*high confidence*). Furthermore, findings suggest that current IPCC emission factors for indirect $N_2O$ continue to underestimate the contribution of this source to total $N_2O$ emissions (*medium confidence*).

#### *2.4.3.4 Future trends and climate change effects*

Climate change is expected to impact $N_2O$ emissions in several ways. Warmer and wetter conditions will enhance the conditions for soil $N_2O$ emissions, acting as a positive feedback to climate change. These conditions have already led to indirect $N_2O$ emissions dominating interannual variability of total emissions (Griffis et al. 2017). Changes in soil moisture driven by changes in precipitation patterns and totals as well as evapotranspiration fluxes will likely dominate the $N_2O$ response to climate change, overshadowing the direct temperature effects on denitrification and nitrification (Fowler et al. 2015). Indeed, changes in precipitation alone are projected to increase total N loading to rivers by 19% within the continental United States by the end of this century, with important implications for indirect $N_2O$ emissions. Offsetting this increase would require a 33% reduction in N application rates (Sinha et al. 2017) A similar dynamic is expected in regions with high N consumption and projected increases in precipitation, such as China, India, and Southeast Asia. However, $N_2O$ emissions are not expected to increase proportionally with N loading, especially as a river becomes N saturated due to an inverse relationship between N loading and removal efficiency, with a doubling in nitrate concentrations estimated to increase $N_2O$ emissions by 40% (Mulholland et al. 2008; Turner et al. 2015) Climate change is also expected to cause changes in land use and management, which will likely impact terrestrial biogeochemical cycles. An increase in the area of irrigated agricultural land could stimulate $N_2O$ emissions increases of 50%-150%, likely a result of increased denitrification activity (Trost et al. 2013; Fowler et al. 2015) .

Analysis by Prentice et al. (2015b) that looked specifically at the land $CO_2$ feedbacks in the CMIP5 and $C^4$MIP models under a range of scenarios indicate that the negative feedbacks associated with increased productivity outweighs the positive feedbacks throughout the 21st century. Arneth et al. (2010) came to a similar conclusion, showing that the feedback range for $CO_2$ fertilisation was between -0.17 to -1.9 $Wm^{-2}$ compared to 0.1 to 0.9 for the positive feedbacks. In both studies however, the inclusion of the nitrogen cycle reduces this $CO_2$ fertilisation effect significantly due to nutrient limitations. Arneth et al. (2010) showed that the negative feedbacks decreased to -0.4 to -0.8 $Wm^{-2}$, which for some models removes the net carbon loss completely. Similarly, when including the feedbacks associated with methane, fire and ozone, the total radiative forcing between the atmosphere and terrestrial biosphere was positive, ranging between 0.9 to 1.5 $Wm^{-2} K^{-1}$ at the end of the 21st Century. Other studies suggest that the cumulative warming effect of methane ($CH_4$) and $N_2O$ emissions over the period 2001-2010 was a factor of two larger than the cooling effect that resulted from $CO_2$ fertilisation (Tian et al. 2016), suggesting that mitigation efforts should be as focused on reducing emissions of these non-$CO_2$ GHGs as on increasing carbon storage capacity. Nevertheless, studies highlight the uncertainty in carbon-cycle estimates from ESMs and the need for more realistic carbon and nutrient cycling in models.

### 2.4.4 Impacts of mitigation on carbon sinks

Under future low emission levels and large negative emissions, the global land and ocean sinks are expected to weaken (or even reverse) (Jones et al. 2016). Carbon today absorbed by the oceans following increases in atmospheric $CO_2$ concentration will partially be released back to the air when concentration declines (Cao and Caldeira, 2010; Ciais et al. 2013; Jones et al. 2016). This means that to maintain atmospheric $CO_2$ and temperature at low levels, both the excess $CO_2$ from the atmosphere and the $CO_2$ progressively outgassed from the ocean and land sinks need to be removed (Cao and Caldeira, 2010). This outgassing from the land and ocean sinks is called the "rebound effect" of the global carbon cycle (Ciais et al. 2013). It will reduce the effectiveness of negative emissions and increase the deployment level needed to achieve a climate stabilisation target (Jackson et al. 2017; Jones et al. 2016).

We calculated the contribution of land use emissions to total anthropogenic emissions over the decade 2003 to 2012 (**Table 2.4**) using several global datasets. The non-CO2 GHG data available in EDGAR do not yet extend beyond 2012. We calculated emissions over the decade to account for inter-annual variability due to ENSO and other sources of variability. We note that land use $CO_2$ emissions values in FAOSTAT, which included emissions from organic soils, are lower than those of both the book keeping models and the DGVMs. To calculate total land use emissions we combined the average emissions from the book keeping models with the average non-$CO_2$ GHG data from the different data sources and we calculated total anthropogenic emissions using CDIAC data for non-land use emissions. Similarly to AR5, we find that global land use emissions are 24% of total anthropogenic emissions (*high confidence*), with slightly more than half these emissions coming as non-$CO_2$ GHGs from agriculture. Since publication of the IPCC Fourth Assessment Report (AR4), land use emissions have remained relatively constant; however, the share of land use in anthropogenic emissions have decreased due to increases in emissions in the energy sector. Increasing non-$CO_2$ emissions in non-Annex I countries may be offsetting decreases in deforestation emissions.

**Table 2.4 Summary of land use fluxes aggregated from 2003 to 2012. Values in bold were used to calculate total-land use emissions**

| Land use emissions | Gtonnes $CO_2$(e) |
|---|---|
| Land use $CO_2$ | |
| **Bookkeeping model average (H&N, Blue)*** | **4.69** |
| DGVM average | 4.75 |
| FAOSTAT | 3.09 |
| Non-$CO_2$ GHGs | |
| Agricultural $CH_4$ | |
| FAOSTAT | 2.78 |
| USEPA | 3.02 |
| EDGAR | 3.72 |
| **Average** | **3.17** |
| Agricultural $N_2O$ | |
| FAOSTAT | 2.17 |
| USEPA | 2.76 |
| EDGAR | 1.82 |
| **Average** | **2.25** |
| **Total emissions (2003 to 2012) from land use** | **10.12** |
| Total anthropogenic emissions (2003 to 2012) from all sources† | 42.07 |
| **Land use emissions : Total emissions** | **24%** |

* Data sources: Hansis et al. 2015; Houghton & Nassikas 2017.
† Total anthropogenic emissions were calculated using CDIAC values for fossil fuel, cement and flaring emissions from the 2016 budget (http://cdiac.ess-dive.lbl.gov/GCP/carbonbudget/2016/) and the land use emissions calculated here.

## 2.5 Historical and future non-GHGs fluxes and precursors of short-lived species from unmanaged and managed land

While the atmospheric concentration of greenhouse gases is the largest factor on anthropogenic changes in climate, the levels of atmospheric aerosol particles (with diameters between about 0.002 µm to about 100 µm), can significantly modulate regional climate and have significant global effects on radiation budget, hydrological cycle and other ecosystem impacts (Boucher et al. 2013; Rogelj et al. 2014; Kok et al. 2018) *(high evidence; high agreement)*. While there was a progress in quantifying regional emissions of anthropogenic and natural land aerosols, considerable uncertainty still remains about their historical trends, their inter-annual and decadal variability and about any changes in the future (Calvo et al. 2013).

Depending on the chemical composition and size, aerosols can absorb or scatter sunlight and thus directly affect the amount of absorbed and scatted radiation (Fuzzi et al. 2015; Nousiainen 2009). In the troposphere, aerosols affect clouds formation and development, and thus change precipitation patterns and amounts (Suni et al. 2015). In addition, deposition of aerosols has implication for surface reflectance, particularly snow, and biogeochemical cycling in critical terrestrial ecosystems nutrients such as nitrogen and phosphorus (Andreae et al. 2002). Primary land atmospheric aerosols are emitted directly into the atmosphere due to natural or anthropogenic processes and include mineral aerosols (or dust), volcanic dust, soot from combustion, organic aerosols from industry, vehicles or biomass burning, bioaerosols, and others. Secondary atmospheric aerosols are particulates that are formed in the atmosphere by gas-to-particles conversion processes from land emissions and accounts for a large fraction of aerosol mass (Hodzic et al. 2016; Manish et al. 2017).

## 2.5.1 Temporal trends, spatial patterns, and variability

### *2.5.1.1 Mineral dust*

One of the most abundant atmospheric aerosols is mineral dust, which is emitted into the atmosphere from arid and semi-arid regions and then transported over long distances across continents and oceans (Ginoux et al. 2001). Depending on the dust mineralogy and size, dust particles can absorb or scatter shortwave and long-wave radiation. Dust particles served as cloud and ice condensation nuclei and influences the optical properties of clouds, their lifetime and consequently, the precipitation rate (Kok et al. 2018). In addition, dust particles have shown to alter the cloud cover through changes in evaporation of cloud droplets (i.e., the cloud burning effect) (Boucher et al. 2013). New and improved understanding of processes controlling emissions and transport of dust, its regional patterns and variability as well as its chemical composition has been developed since AR5.

Characterisation of spatial and temporal distribution of dust emissions is essential for weather prediction and climate projections (*robust evidence, high agreement*). Satellite observations have been the most effective way of identifying and quantifying regional dust sources, which were initially derived by analysing Aerosol Index (AI) or Aerosol Optical Thickness (AOT) (e.g., from sensors such as the Total Ozone Mapping Spectrometer (TOMS) the Moderate Resolution Imaging Spectroradiometer (MODIS), the Ozone Monitoring Instrument (OMI), the Multi-angle Imaging Spectroradiometer (MISR) (Lyapustin et al. 2018; Kocha et al. 2013).

The new 'dust source activation' (DSA) frequency method was developed and applied over the Sahara desert by analysing output from the Spinning and Enhanced Visible and Infrared Imager (SEVIRI) sensor on the geostationary Meteosat Second Generation (MSG) satellites (Ashpole and Washington 2013; Ian and Richard 2013) ). Characterisation of dust emission further improved with analysis of aerosol vertical profiles from the Cloud-Aerosol Lidar with Orthogonal Polarization (CALIOP) instrument on board the Cloud-Aerosol Lidar and Infrared Pathfinder Satellite Observation (CALIPSO) satellite (Todd and Cavazos-Guerra 2016). The Dust Emission Index derived from the CALIOP time series shows that the highest emissions over Sahara occur during June-September, with the night-time emission, driven by convective cold pools, exceeding the day-time emissions. Although there is a growing confidence in characterising the seasonality and peak dust emissions (i.e., spring-summer, (Wang et al. 2015) and how the meteorological and soil conditions control dust sources, an understanding of long-term future dust dynamics, inter-annual dust variability and how they will affect future climate still requires substantial work.

While satellites remain the primary source of information about the dust distribution in the atmosphere and are used to derive emissions of dust, new surface observations improve process understanding and reduce uncertainty about optical and mineralogical properties of the dust (Rocha-Lima et al. 2018). Dust particles include several minerals with different physical and chemical properties, which affect how dust interacts with radiation and hydrological cycle. Mineralogy of dust is strongly linked to mineralogy of soils where dust is emitted. New global databases were developed to characterise mineralogical composition of soils for use in the weather and climate models (Journet et al. 2014; Perlwitz et al. 2015) New field campaigns as well as new analysis from prior campaign have produce accurate characterisation of optical properties and insights into role of dust in climate system: i) the SHADOW (study of SaHAran Dust Over West Africa) campaign in Mbour, Senegal (14◦ N, 17◦ W) in March–April 2015 (Veselovskii et al. 2016); ii) the SALTRACE (Saharan Aerosol Long-range Transport and Aerosol-Cloud interaction Experiment) in Barbados, June and July 2013, to characterise long-range transported transport of Saharan Dust across the Atlantic Ocean (Groß et al. 2015), ii) the UK Ice in dust clouds experiment took place in Cape Verde, August of 2015, to characterise aerosol particles and their ability to act as ice nuclei (IN) and cloud condensation nuclei (CCN) within convective and layered clouds (Price et al. 2018).

### *2.5.1.2 Carbonaceous Aerosols*

Carbonaceous aerosols are one of the most abundant components of particles in the continental areas of the global atmosphere (Contini et al. 2018). It can comprise about 60–80% of PM1 (Particulate Matter with size less than 1 micrometer) in urban and remote atmosphere (Tsigaridis et al. 2014; Kulmala and Asmi 2011). It comprises of an organic fraction (Organic Carbon - OC) and a refractory light absorbing component, generally referred as Elemental Carbon (EC) (Gilardoni et al. 2011). OC is a major component of aerosol mass concentration, and it originates from different anthropogenic (combustion processes) and natural (biogenic emissions) sources (Robinson et al. 2007). A large fraction of OC in the atmosphere has a secondary origin, since OC can be both primarily emitted but also formed in the atmosphere through condensation to the aerosol phase of low vapour pressure compounds emitted as primary pollutants or formed in the atmosphere. This component is called Secondary Organic Aerosol (SOA) (Hodzic et al. 2016). Organic carbon is also characterised by a high solubility with a high fraction of water soluble organic aerosol (WSOA) and it is one of the main drivers of the oxidative potential of atmospheric particles. This makes OC an efficient CCN in most of the conditions (Pöhlker et al. 2016; Thalman et al. 2017; Schmale et al. 2018a). In terms of radiative effects and optical properties, OC is important for the scattering properties of aerosols and EC is important for the absorption component (Tsigaridis et al. 2014; Fuzzi et al. 2015; Rizzo et al. 2013). A third components is the so-called brown carbon (BrC) that has assumed an increasing importance because this organic material shows enhanced absorption at short wavelengths (Wang et al. 2016d; Laskin et al. 2015; Liu et al. 2016b; Bond et al. 2013; Pöhlker et al. 2016; Saturno et al. 2018).

Biomass burning is a major global source of carbonaceous aerosols (Bowman et al. 2011; Harrison et al. 2010; Reddington et al. 2016; Artaxo et al. 2013). As knowledge of past fire dynamics improved through new satellite observations, new fire proxies’ datasets (Marlon et al. 2013; van Marle et al. 2017) , and process-based models (Hantson et al. 2016), a new historic biomass burning emissions dataset starting in 1750 has been developed (Van Marle et al. 2017). Revised versions of OC emissions (Van Marle et al. 2017) show in general smaller trends than emissions developed by (Lamarque et al. 2010) for CMIP5.

Previous emission inventories of BC were mostly obtained in a bottom-up framework (Jacobson 2012; Wang et al. 2014a)), an approach that derives emissions based on categorised emitting sources and emission factors used to convert burning mass to emissions. (Bond et al. 2013) estimated that at pre-industrial time (around 1750s) emission from biofuel and biomass burning were approximately 1400 Gg of black carbon per year. Although it’s not clear how much of such discrepancy is related to the underestimates of biomass burning BC. (Wang et al. 2014a) estimated a substantial contribution of agricultural fires and wildfires to BC emissions for 1960 to 2007 period, as can be observed from Figure 2.17.

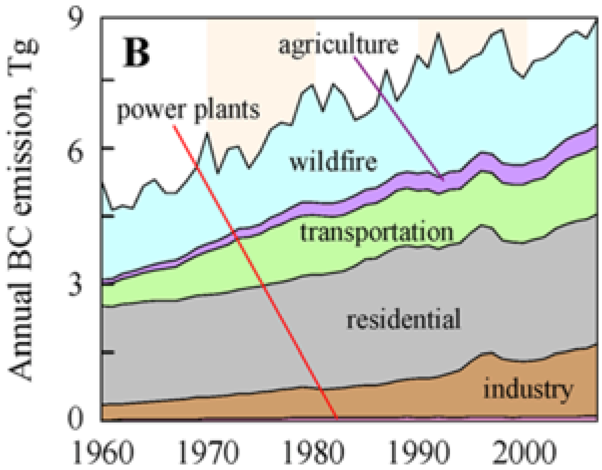


**Figure 2.17 Global temporal trends of annual BC emissions (B) in various sectors, from (Wang et al. 2014a)**

A top-down total (biomass burning, fossil fuel consumption, etc.) global estimate of BC emission17.8 ± 5.6 Tg $yr^{-1}$ was obtained using a fully coupled climate-aerosol-urban model constrained by aerosol absorption optical depth and surface concentrations from global and regional networks (Cohen and Wang 2014). Cohen and Wang 2014 estimate is a factor of 2 higher than previous estimates (e.g. 7.6–8.8 Tg $yr^{-1}$, (Bond et al. 2007; van der Werf et al. 2010)), with considerably higher BC emissions for Eastern Europe, Southern East Asia, and Southeast Asia mostly due to higher anthropogenic BC emissions estimates. (Giglio et al. 2013) found a gradual fire area decrease of 1.7 Mha $yr^{-1}$ (–1.4% $yr^{-1}$) in Northern Hemisphere Africa since 2000, a gradual increase of 2.3 Mha $yr^{-1}$ (+1.8% $yr^{-1}$) in Southern Hemisphere Africa also since 2000, a slight increase of 0.2 Mha $yr^{-1}$ (+2.5% $yr^{-1}$) in Southeast Asia since 1997, and a rapid decrease of approximately 5.5 Mha $yr^{-1}$ (-10.7% $yr^{-1}$) from 2001 through 2011 in Australia, followed by a major upsurge in 2011 that exceeded the annual area burned in at least the previous 14 years. The net trend in global burned area from 2000 to 2012 was a modest decrease of 4.3 Mha $yr^{-1}$ (-1.2% $yr^{-1}$). In the Brazilian Amazonia, deforestation was reduced from 27,000 $Km^2$ in 2004 to 5,000 $Km^2$ in 2012 (Artaxo et al. 2013), but in recent years have increased to 7,500 $Km^2$ in 2017.

#### *2.5.1.3 Biogenic Volatile Organic compounds (BVOCs)*

Land use change can affect the climate through changed emissions of short lived climate forcers (SLCFs) such as aerosols, ozone and methane. Biogenic volatile organic compounds (BVOCs) are emitted in large amounts by broadleaf forests while crops and grasslands emit very little or no BVOC (Guenther et al. 2012). BVOCs are rapidly oxidized in the atmosphere to form less volatile compounds that can condense and form secondary organic aerosol (SOA). This can affect the aerosol size distribution both by contributing to new particle formation and by growth of larger pre-existing particles. This affects the scattering of radiation by the particles themselves (direct aerosol effect), but also changes the amount of cloud condensation nuclei and the optical properties of clouds (indirect aerosol effect). These BVOCs includes isoprene, terpenes, alkanes, alkenes, alcohols, esters, carbonyls and acids (Peñuelas and Staudt 2010; Guenther et al. 1995, 2012). These BVOCs emissions represent a carbon loss to the ecosystem, where their emission can represent up to 10% of the carbon fixed by photosynthesis under stressful conditions (Bracho-Nunez et al. 2011). The global average emission for vegetated surfaces is 0.7g C $m^{-2}$ $yr^{-1}$ but could exceed 100 g $m^{-2}$ per year in some tropical ecosystems (Peñuelas and Llusià 2003). For some of the compounds, BVOC emissions depend on temperature, and in a warming planet, it is expected that emission rates of most BVOCs will increase (Peñuelas and Llusià 2003). It does so not only by enhancing the enzymatic activities of synthesis but also by raising the BVOCs vapour pressure and by decreasing the resistance of the diffusion pathway. BVOC emissions are thus expected to increase sharply as global temperatures rise *(moderate evidence, high agreement).*

By applying the most frequently used algorithms of emission response to temperature, it can be estimated that climate warming over the past 30 years have already increased BVOC global emissions by 10% since the preindustrial times (*limited evidence and medium agreement*). A further 2 ºC –3ºC rise in the mean global temperature, which possibly could occur during this century (Michelman et al. 1991) could increase BVOC global emissions by an additional 30–45% (Peñuelas and Llusià 2003). Furthermore, global warming in boreal and temperate environments not only means warmer average and warmer winter temperatures but also implies an extended plant activity season (Peñuelas 2009) increasing total annual emissions even further. There is, moreover, a lack of precise and complete data on the effects of all the other global change components such as land use changes or global fertilisation with increasing $CO_2$ and N inputs, but everything seems to indicate that the most likely overall effect will be to increase BVOC emissions (Peñuelas and Staudt 2010).

Changing BVOC emissions also affect the oxidant concentrations in the atmosphere. The impact on the concentration of ozone depends on the local NOx concentrations. In more polluted regions, higher BVOC emissions lead to increased production of ozone. Enhanced ozone leads to formation of more OH and a reduction in methane lifetime. In more pristine regions (NOx-limited), increasing BVOC emissions instead leads to decreasing OH and ozone concentrations, resulting in a longer methane lifetime. The net effect of BVOCs must be quantified by models, and can change over time if NOx emissions (from all sources) are changing.

BVOC are the most important precursors of secondary organic aerosols, via oxidation and chemical processes. The high amount of SOA over boreal and tropical forests are mostly originated from BVOC emissions (Manish et al. 2017). Recent findings show that the emissions of BVOCs can be the starting point of aerosol formation in the upper atmosphere in Amazonia, where they produce particles that are oxidized, reducing their volatility and transported back to lower atmosphere where it populates the lower atmosphere (Schulz et al. 2018; Wang et al. 2016a; Andreae et al. 2018). These organic particles produced in the upper troposphere could be the main CCN population in Amazonia that are responsible for the vigorous hydrological cycle (Pöhlker et al. 2018).

Over forest areas, most of aerosol particles are organic with traces of inorganic components that makes them efficient CCN, making a strong link between BVOC emissions by plants and climate/hydrological cycle (Fuentes et al. 2000; Schmale et al. 2018b; Pöhlker et al. 2016, 2018). Many studies have clearly shown that BVOC emissions are highly sensitive to land use change, climate change, and other disturbances but we need improvements on the processes controlling the specific responses (Jardine et al. 2011, 2015; Fuentes et al. 2016).

Rising temperature and $CO_2$, when it enhances photosynthesis, can increase BVOC emissions and thus contribute to dampen the initial warming (Arneth et al. 2010; Kulmala et al. 2003). This negative feedback occurs through the response to larger amounts of secondary organic aerosols in the atmosphere (Section 2.6.1). The scattering of solar radiation by those aerosol particles will also increase the ratio of diffuse to direct radiation, which can boost gross primary production suggesting an enhanced negative feedback (Kulmala et al. 2014).

There is *limited evidence and medium agreement* that the decrease in the emissions of biogenic volatile organic compounds (BVOC) resulting from the historical conversion of natural vegetation, in particular forests, to cropland has impacted climate through atmospheric chemistry (Section 2.5). Lower BVOC emissions cause a decrease in the global formation of secondary organic aerosols (-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 2014a) through the direct aerosol effect. The large simulated range is explained by the remaining uncertainty in both the change in BVOC emissions and the SOA production in global models. In present-day conditions, global SOA production from all sources spans between 13 and 121 Tg $yr^{-1}$ (Tsigaridis et al. 2014). The indirect aerosol effect (change in cloud condensation nuclei), resulting from land use induced changes in BVOC emissions, add an additional positive radiative forcing of 0.008 W $m^{-2}$

by(Scott et al. 2017). This is only one estimate and more studies with different model setups are needed to fully assess the indirect aerosol effect associated with land use change from the preindustrial to present.

There is also *limited evidence* that historical changes in BVOC emissions have also impacted tropospheric ozone *(medium agreement).* 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 these $NO_x$ limited regions (Heald and Geddes 2016). Globally, land use change effects on BVOC emissions between the pre-industrial and present day have decreased the tropospheric ozone concentrations, and thus led to negative radiative forcing with values from 3 studies ranging from -0.01 to -0.13 W $m^{-2}$ (Scott et al. 2017; Heald and Geddes 2016; Unger 2014a). The spread in the forcing can again be partly explained by different magnitudes of the change in BVOC emissions between the pre-industrial and present day.

A third impact of historical decrease in BVOC emissions is the reduced 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 2014a). 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 (Xu et al. 2008; Scott et al. 2017).

Lastly, 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.

Since industrial revolution, emissions from all major non-GHG have increased significantly (Myhre et al. 2013; Lamarque et al. 2010) as can be observed in Figure 2.18. BC in particular have increase 5 times in terms of emissions, which gives BC a strong positive radiative forcing, that is partially counterbalanced by the negative forcing of OC emissions (Bond et al. 2013).

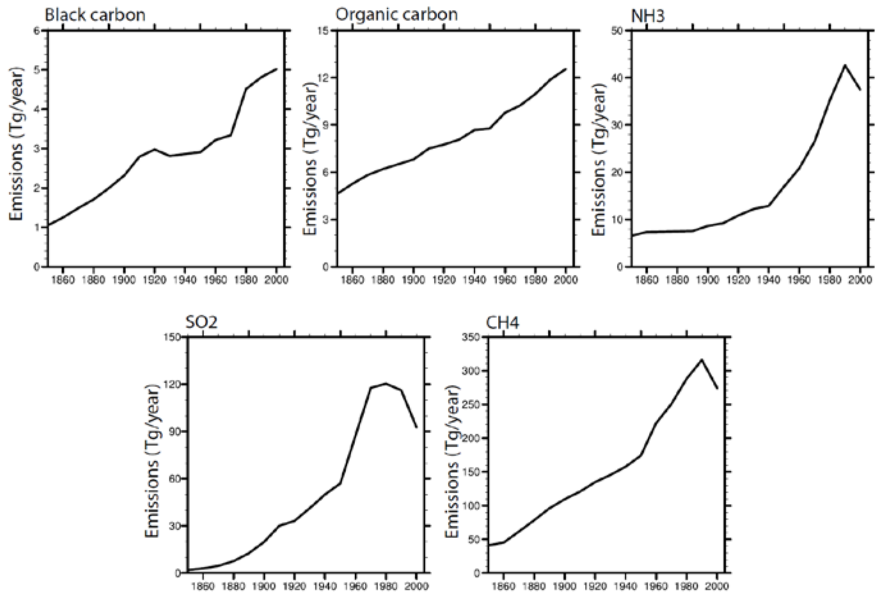


**Figure 2.18 Time evolution from 1850 of the land anthropogenic emissions for black carbon (Tg(C)/year), organic carbon (Tg(C)/year), ammonia (Tg(NH3) yr$^{-1}$), sulfur dioxide (Tg(SO2) yr$^{-1}$), and methane (Tg(CH4) yr-$^{1}$) (Lamarque et al. 2010)**

### *2.5.1.4 Impacts on climate of future global land use scenarios*

Anthropogenic land use change, rising $CO_2$ levels and climate change will all affect future BVOC emissions. There is only a very limited number of studies investigating the climate impacts of BVOCs using future land use scenarios. 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 W m$^{-2}$ (decreased reflecting particles in the atmosphere) and a first indirect aerosol forcing of -0.001 W m$^{-2}$ (change in the amount of cloud condensation nuclei). 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. 2012b; 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.

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 W m$^{-2}$ K (Scott et al. 2018b; Paasonen et al. 2013). These studies include only aerosol impacts on cloud droplet number concentration, while a modelling study using a more interactive cloud schemes also found BVOC effects on cloud water path and cloud fraction (Sporre et al. 2018). The feedback could hence be larger if also more aerosol-cloud interactions are considered. Enhanced aerosol scattering from increasing BVOC emissions has been estimated to contribute with a global gain in BVOC emissions of 1.07 (Rap et al. 2018).

Future ozone concentrations will also be affected by land use strategies and how these affect emissions of BVOCs and $NO_x$ from the soil. Investigations of only land use changes (according to RCP8.5) on future ozone concentrations found the radiative forcing from the decrease in ozone concentrations to be quite small (-0.002 W m$^{-2}$) (Scott et al. 2018a). The radiative forcing from changing methane lifetime was of a similar magnitude at -0.003 W m$^{-2}$. None of the studies investigating future changes in BVOC emissions due to $CO_2$ and climatic effects has quantified the impact on climate through ozone and methane. Nevertheless, from process studies it is likely that increasing BVOC emissions in the future could lead to, at least on a global scale, increasing ozone production and a longer methane lifetime creating positive climate feedbacks. The aerosol effects associated with feedbacks through BVOC emissions are, at least globally, opposite of those from ozone and methane, and it is possible that the different feedbacks will partly cancel each other out.

### 2.5.2 Dust, BVOC and carbonaceous aerosols in Coupled Climate and Earth System Models

Coupled Model Intercomparison Project, phase 5 (Taylor et al. 2012; Dirmeyer et al. 2013) included a number of models with representation of aerosol emission, transport, and deposition. Such models either specify emissions of short-lived gases into their atmospheric components or had prognostic capabilities for some precursors such as dust. Since CMIP5 a number of coupled and atmospheric modelling studies included prognostic emissions for dust, carbonaceous aerosols, and BVOC. Earth System Models (ESM) have difficulties in properly model BVOCs and SOA production. Actually all CMIP5-class ESM did not include explicitly SOA formation, due to the chemical complexity and diversity of process that depends heavily on land use (Arneth et al. 2011). BVOC emissions are very sensitive to temperature and radiation fields, and models such as MEGAN version 2.1 (Model of Emissions of Gases and Aerosols from Nature) (Guenther et al. 2012) been incorporated in the Community Land Model version 4 (CLM4), but they are too computationally intensive to be included in ESM. MEGAN take into account about 150 specific compounds that are still a fraction of ecosystem and climatic relevant BVOCs (Isaacman-Vanwertz et al. 2018; Park et al. 2013).

Analysis of the results from 23 CMIP5 models reveals that all models systematically under-estimate dust emissions, amount of dust in the atmosphere and its inter-annual variability (Evan et al. 2014). The vertically integrated mass of atmospheric dust per unit area (i.e., mean dust mass path DMP, g $m^{-2}$) , obtained from the advanced very high resolution radiometer (AVHRR) (Stowe et al. 2002) for 1982–2004 and the Moderate Resolution Imaging Spectroradiometer (MODIS) (Remer et al. 2005) Terra instrument for 2000–2013 was approximately 0.6–1.0 g $m^{-2}$, while the 23 CMIP models range for DMP was only 0.05– 0.46 g $m^{-2}$ (Evan et al. 2014). The relationship between CMIP5 multi-model DMP and total northern Africa emissions implies that the AVHRR- and MODIS-based dust emissions are 3 times as high as those used in the models, that is $4500 \pm 1500$ Tg $yr^{-1}$ (Evan et al. 2014). General circulation models (GCMs) typically do not reproduce inter-annual and longer time scales variability seen in observations (Evan et al. 2016).

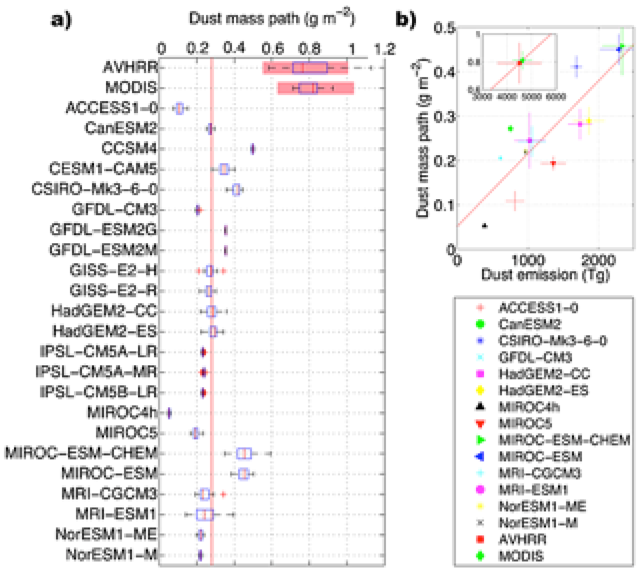


**Figure 2.19 Comparison of modelled and remotely sensed dust mass paths and emissions, from (Evan et al. 2014)**

### 2.5.3 Contribution of non-GHG fluxes from managed and unmanaged lands to atmospheric composition

#### *2.5.3.1 Mineral dust*

Usually mineral dust is considered as a "natural" aerosol as it is produced by wind over dry regions with low-density vegetation. Soil and vegetation cover could be altered by land use land cover changes or agricultural practices. (Stanelle et al. 2014) used a global climate-aerosol model and found that global annual dust emissions have increased by 25% from preindustrial to present day (e.g., from 729 Tg $yr^{-1}$ to 912 Tg $yr^{-1}$) with 56% increase driven by climate change and 40% by land use cover change such as conversion of natural lands to agriculture. Approximately 10% of present day dust emissions originate from agricultural regions.

In North Africa most dust is of natural origin with the recent 15% increase in dust emissions attributed to climate change. In North America two thirds of dust emissions take place on agricultural lands and both climate change and land use change jointly drive the increase. Between pre-industrial and present-day the overall effect of changes in dust is - 0.14 $Wm^{-2}$ cooling of clear sky net radiative forcing on top of the atmosphere, with -0.05 W $m^{-2}$ form land use and -0.083 W $m^{-2}$ from changes in climate.

The observed decreasing trends in Sahel dust emissions and transport has been attributed to reduction in surface winds primarily due to increased vegetation surface roughness ("stilling" effect) with secondary effects form changes in turbulence and evapotranspiration, and changes large-scale circulation (Cowie et al. 2013). Similarly, the observed decreasing trends in dust storms in Northeast Asia since the 1950s, with the exception of beginning of the 21st century, has been attributed to surface wind stilling. In addition, analysis of relationship of vegetation green-up dates derived from the satellite observation and dust storms from 1982 to 2008 over Inner Mongolia, Northern China showed a significant dampening effect of earlier spring greening on dust storms ($r = 0.49, p = 0.01$), with a one-day earlier green-up date corresponding to a decrease in annual spring dust storm outbreaks by 3%.

One commonly suggested reason for the lack of dust variability in climate models is the models' inability to simulate the effects of land surface changes on dust emission (Stanelle et al. 2014). There has been progress in incorporating effects of vegetation, soil moisture, surface wind and vegetation on dust emission source functions and show more agreement with the satellite observations both in terms of AOD and DMP (Kok et al. 2014). New prognostic dust emissions models now able to account for both changes in surface winds and vegetation characteristics (e.g., leaf area index and steam area index) and soil water, ice, and snow cover (Evans et al. 2016). As a result, the new modelling studies (e.g. Evans et al. 2016) indicate that in regions where soil and vegetation respond strongly to ENSO events, such as in Australia, inclusion of dynamic vegetation characteristics into dust emission parameterisations improves comparisons between the modelled and observed relationship long-term climate variability (e.g., ENSO) and dust levels (Evans et al. 2016).

While inter-annual climate variability, particularly precipitation, often is the primary driver of regional dust variability, changes in the dust radiative forcing induce climate feedbacks that affect regional precipitation, as it been illustrated for the summer precipitation during the 2000–2009 over southern India (Solmon et al. 2015).

#### *2.5.3.2 Carbonaceous aerosols*

Carbonaceous aerosols are important in urban areas as well as pristine continental regions. In boreal and tropical forests, OC originates from BVOC oxidation, being isoprene and terpenes the most important precursors (Claeys et al. 2004; Hu et al. 2015; De Sá et al. 2017, 2018; Liu et al. 2016c) In particular, isoprene epoxydiol-derived secondary organic aerosol (IEPOX-SOA) is being identified in recent studies in North America and Amazonian forest as a major component in the oxidation of isoprene (Allan et al. 2014; Schulz et al. 2018; De Sá et al. 2017). The largest global source of BC aerosols is open burning of forests, savannah and agricultural lands with about 2,700 Gg $yr^{-1}$ in the year 2000 (Bond et al. 2013).

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

One important component of carbonaceous aerosols are the primary biological aerosol particles (PBAP), also called bioaerosols, that correspond to a significant fraction of aerosols in forested areas (Fröhlich-Nowoisky et al. 2016; Pöschl and Shiraiwa 2015). They are emitted directly by the vegetation as part of the biological processes (Huffman et al. 2012). Airborne bacteria, fungal spores, pollen, archea, algae, and other bioparticles are essential for the reproduction and spread of organisms across various terrestrial ecosystems. They can serve as nuclei for cloud droplets, ice crystals, and precipitation, thus influencing the hydrological cycle and climate (Whitehead et al. 2016; Scott et al. 2015; Pöschl et al. 2010).

#### *2.5.3.3 BVOCs*

BVOCs' possible climate effects have received little attention because it was thought that the short lifetime of BVOCs would preclude them from having any significant direct influence on climate (Unger 2014b; Sporre et al. 2018). Higher temperatures and increased $CO_2$ concentrations are (separately) expected to increase the emissions of BVOCs. This has been proposed to initiate negative climate feedback mechanisms through increased formation of SOA. More SOA can make the clouds more reflective, which can provide a cooling. Furthermore, the increase in SOA formation has also been proposed to lead to increased aerosol scattering, resulting in an increase in diffuse radiation. This could boost gross primary production (GPP) and further increase BVOC emissions. These important feedbacks are starting to emerge (Sporre et al. 2018; Kulmala et al. 2003; Arneth et al. 2017) 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 *(little evidence, moderate agreement)*. Either directly, by reflecting more solar radiation, or indirectly, by increasing CCN, the increase in scattering aerosols reduces the amount of solar radiation reaching the surface of Earth with a consequent cooling effect. (Goldstein et al. 2009) observed aerosol optical thickness resulting from BVOCs which in summer is sufficient to form a regional cooling haze over the South-eastern USA (i.e., it constitutes a significant potential for a regional negative feedback on climate warming). Furthermore, aerosols scatter the light received by the forest canopy, increasing $CO_2$ fixation (Niyogi et al. 2004) and providing another indirect, potentially negative feedback on warming. As a result, there should be a net cooling of the Earth's surface during the day. Most tropical forest BVOC are primarily emitted from foliage of trees but soil microbes can be a major source of some compounds including sesquiterpenes (Bourtsoukidis et al. 2018).

Both tropospheric ozone and methane are potent greenhouse gases and changing BVOC emissions can affect climate through their effects on these gases. In a future scenario with increasing BVOC emissions, higher ozone and methane concentrations could lead to an enhanced warming which could further increase BVOC emissions and contribute with a positive climate feedback (Arneth et al. 2010). This feedback is a non-linear function of the NOx levels and is therefore dependent on future NOx emissions.

The processes associated with BVOC emissions, oxidation, aerosol formation and their climatic impact are still associated with large uncertainties, and thus point to a need for further developing scientific research. However, it has also been observed that BVOCs help to slow down nocturnal cooling in areas with relatively dry air masses and active photosynthesis (Hayden 1998). Makkonen et al. 2012 used the global aerosol-climate model ECHAM5.5-HAM2 to explore the effect of BVOC emissions on new particle formation, clouds and climate and found that the change of shortwave cloud forcing from year 1750 to 2000 ranged from -1.4 to -1.8 $Wm^{-2}$, and that from the year 2000 to 2100 ranged from 1.0 to 1.5 W $m^{-2}$.

Apart from the direct local BVOC greenhouse effect, which has detectable effects only when canopy-scale BVOC emissions are high, an additional global indirect greenhouse effect must also be considered because BVOCs increase the production of ozone and change the atmospheric lifetime of methane, and hence enhance the greenhouse effect of these other gases. It therefore seems that the increases in BVOC emissions expected as a result of the current warming and global changes could thus significantly contribute (via negative and positive feedbacks) to the complex processes associated with global warming. Whether the increased BVOC emissions will cool or warm the climate depends on the relative weights of the negative (increased albedo and $CO_2$ fixation) and positive (increased greenhouse action) feedbacks (Peñuelas and

Llusià 2003). The net chemical forcing of global climate due to all known anthropogenic influences on BVOC emissions has been estimated in -0.17 W $m^{-2}$ (cooling). This magnitude of forcing is comparable to that of the surface albedo. Many questions about these BVOC relationships with climate remain to be solved, so laboratory experiments, global climate modelling and extensive international measurement campaigns are necessary to quantify BVOC emissions, aerosol formation and reactions with hydroxyl radicals and ozone, among other processes including soil or water deposition. The still scarce available data indicate though that BVOC emissions should be included in assessments of anthropogenic radiative forcing.

How BVOCs will affect climate in the future is highly dependent on what will happen to the emissions, which is still uncertain. One recent study including dynamic vegetation, land use change, $CO_2$ 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). Nevertheless, regionally and locally, the changes in the BVOC emissions can still be substantial. The number of studies investigating BVOC impacts on climate is still limited. The existing studies indicate that changing BVOC emissions have the potential to affect climate and even in studies where global forcing numbers are small, the regional or local impacts of changing BVOC emissions can be important. However, there is a still 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 a is need for continued research into these processes but the current knowledge indicates that changing BVOC emissions should be taken into consideration when assessing the future climate and how land use will affect it.

### 2.5.4 Changes in hydrological cycle

Aerosol-cloud interactions remain the greatest uncertainty in our understanding of anthropogenic perturbations to the climate system from the point of view of precipitation and also from the radiative balance (Boucher et al. 2013; Fan et al. 2012; Rosenfeld et al. 2014; Rosenfeld 2000). The ability of atmospheric particles to act as seeds for cloud droplets provides the most effective means for them to influence terrestrial ecosystems through precipitation (Carslaw et al. 2013; Spracklen et al. 2008). The changes in CCN population and properties is responsible for the cloud adjustments that are key to the so-called aerosol indirect effects (McFiggans et al. 2006; Stevens and Feingold 2009). The net aerosol effective radiative forcing from aerosol-cloud interactions is negative, so anthropogenic changes in clouds cool the climate by about -0.55 W $m^{-2}$, with a large uncertainty ((Boucher et al. 2013).

Aerosols can both enhance or suppress precipitation, depending on atmospheric thermodynamic conditions, water vapor availability, cloud phase (warm or ice) and CCN amounts and properties (Koren et al. 2014, 2012, 2007; Silvestrini et al. 2011; Camponogara et al. 2018; Machado et al. 2018; Andreae et al. 2018; Cirino et al. 2014). These complex relationships are very non-linear and interdependent. There exists *robust evidence* that forest fires and urban aerosols increase CCN and the number of activated cloud droplets (Andreae et al. 2018, 2004; Freud et al. 2008) For the assumption of constant liquid water content, increased droplet numbers will directly translate into a reduction of cloud droplet radii, increasing cloud albedo and exerting a negative (cooling) radiative effect on the top-of-atmosphere radiation balance. It should be noted that the uncertainties associated with mixed- and ice-cloud microphysics remain significant. The current global and regional climate models have important limitations in the parameterisation of convective clouds, do not normally include convective cloud microphysics and hence lack the ability to represent the majority of the effects proposed to be of importance for the aerosol effects on clouds and precipitation. However, both recent development of advanced aerosol-aware convection parameterisations as well as the increasing availability of (near) global cloud resolving modelling will help to close this gap in the medium to long-term.

Recent studies on freshwater availability shows that it is changing worldwide for several different reasons. Groundwater, soil moisture, surface waters, snow and ice are all dynamic components of the terrestrial water cycle. Recent studies have identified locations where terrestrial water storage (TWS; the sum of these five components). Rodell et al. (2018) have quantified observed trends in terrestrial water storage measured by the Gravity Recovery and Climate Experiment (GRACE) satellites during 2002–2016. They categorise their drivers as natural interannual variability, unsustainable groundwater consumption, climate change or combinations thereof. Some of these changes are consistent with climate model predictions. This observation-based assessment of how the world's water landscape is responding to human impacts and climate variations provides a blueprint for evaluating and predicting emerging threats to water and food security in terrestrial ecosystems.

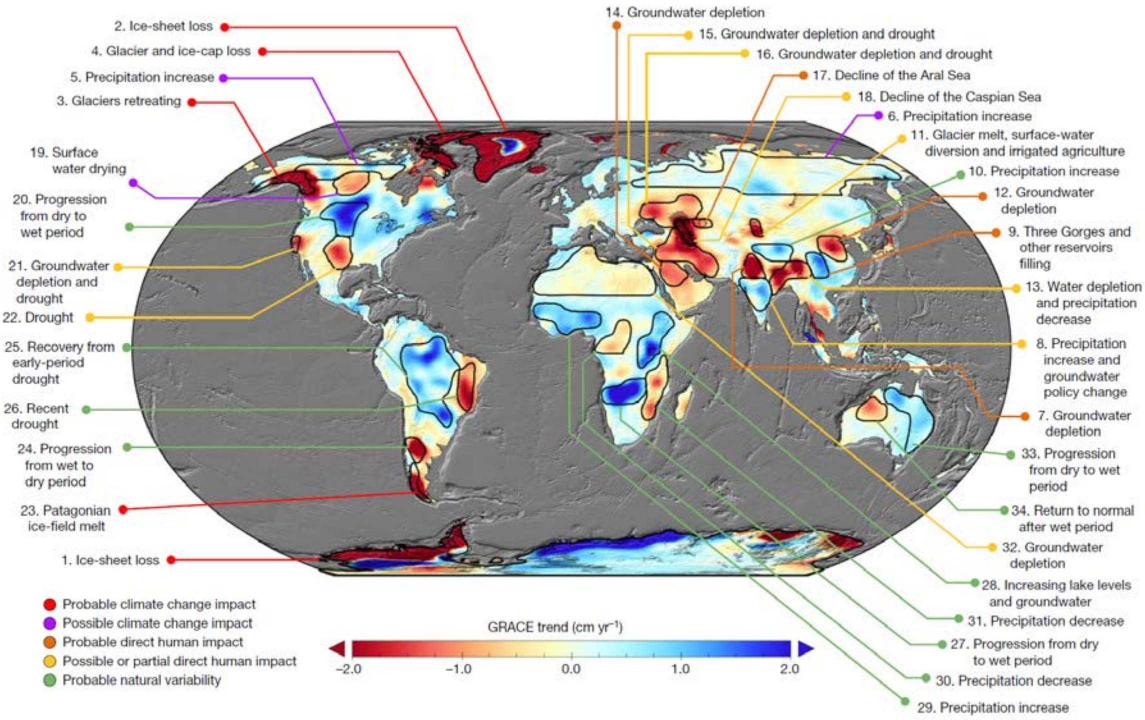


**Figure 2.20 Annotated map of terrestrial water storage (TWS) trends. Trends in TWS (in centimetres per year) obtained on the basis of GRACE observations from April 2002 to March 2016. The cause of the trend in each outlined study region is briefly explained and colour-coded by category (Rodell et al. 2018)**

## 2.6 Land-induced changes on climate and weather, via both biophysical effects and changes in net CO2 emissions from land

The evidences 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 as further discussed below.

The studies that have examined the respective roles of oceans and land on the initiation of the last glaciation, that occurred 115000 years ago, concluded that vegetation feedbacks played a major role during that specific climatic transition. They agreed that glacial inception would not have been possible without the existence of feedbacks from changes in land-cover distribution (De Noblet et al. 1996; Kageyama et al. 2004). 6000 years ago, during the mid-Holocene, the Sahara was greener than today and according to climate models this greening contributed to maintain a rather intense African monsoon (De Noblet-Ducoudré et al. 2000).

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. He hypothesised that the observed reduction in rainfall in the 1970s may be linked to overgrazing north of 18°N in northern West Africa. The grazing-induced increase in land surface albedo indeed resulted in increased radiative cooling of the atmosphere and compensating sinking motion. This sinking motion tends to suppress rainfall and thus sustain the desertified system.

Since then there have been many modeling works that reported impacts of idealised or simplified land cover changes on weather patterns (Pielke et al. 2011). The number of publications dealing with such issues increased significantly over the past 10 years, with more studies that address realistic past or projected land changes. The fraction of those papers that addresses the impacts of changes in land management remains however quite low 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. 2015c; Duveiller et al. 2018b; Forzieri et al. 2017) and the literature is therefore limited.

The focus of this section is summarised Figure 2.21. We will 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 $CO_2$ from land. Land-induced changes in atmospheric chemistry, aerosols, etc. are discussed in Section 2.5.

All sections discuss impacts of land on global and regional climate, and climate extremes, whenever the information is available. Section 2.6.1 presents effects of historical and future scenarios; Section 2.6.2 is devoted to impacts of specific anthropogenic land uses such as afforestation, deforestation, irrigation, urbanisation; Section 2.6.3 focuses on how climate driven land changes feedback on climate and Section 2.6.4 puts forward that land use changes in one region can affect another region.

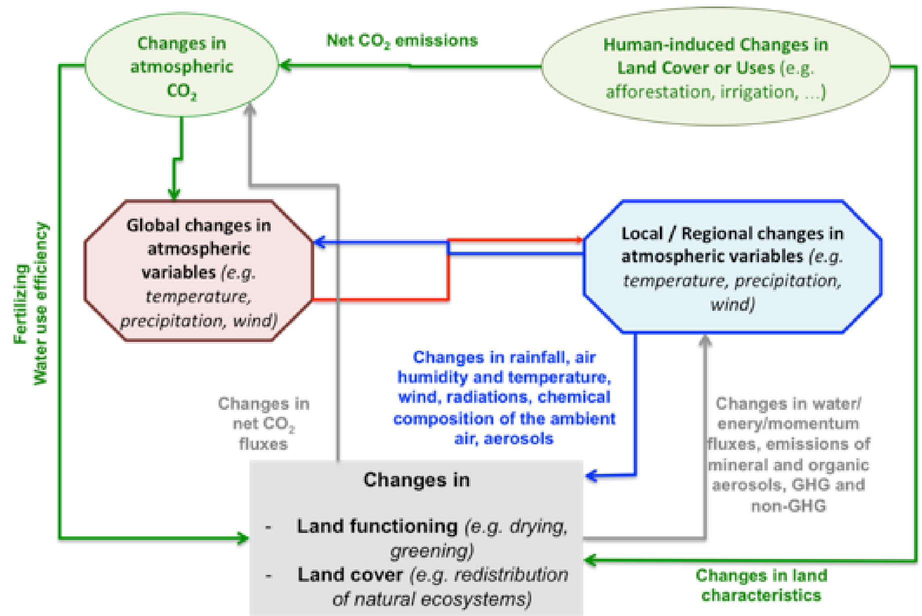


**Figure 2.21 Global, local and regional climate changes are the focus of this section. They are examined via changes in climate states (e.g., changes in air temperature and humidity, rainfall, radiations) as well as via changes in atmospheric dynamics (e.g., circulation patterns). Changes in land that influence climate are either climate- or Human- driven. Green arrows and boxes refer to what we consider herein as imposed changes. 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**

### 2.6.1 Impacts of Historical and Future IPCC land-use scenarios

The studies reported below focus essentially on modeling 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. Each section starts with describing changes at the global scale, then 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.

#### *2.6.1.1 Impacts of global historical land use changes on climate*

##### *2.6.1.1.1 At the global level*

There is *limited evidence and no agreement* that historical land use changes have contributed to net global warming throughout the 20th century. Evidences are available from only four modeling studies. Change in global annual surface temperature ranges from -0,05°C between years 1850 and 2000 (Brovkin et al. 2004), to 0.13°C –0.15°C for the 20$^{th}$ century (Pongratz et al. 2010). A net global warming of 0.15°C is also found by (Matthews et al. 2004) when considering land use changes since 1700, while a net cooling of -0.02°C is found by (Simmons and Matthews 2016) during the same time period.

Despite this net 'no change' signal, there is *robust evidence and high agreement* that the biophysical effects alone of those land use changes had a cooling effect on the global mean temperature (Pongratz et al. 2010; Brovkin et al. 2004; Matthews et al. 2004; Strengers et al. 2010; Simmons and Matthews 2016; De Noblet-Ducoudré et al. 2012). The estimated global simulated biophysical cooling is -0.10 ± 0.05°C and modelled responses range from -0.57°C to +0.06°C (Table 2.5). This cooling is essentially dominated by changes in surface albedo as historical land use changes have led to a dominant brightening of land as discussed in AR5 (Myhre et al. 2013). Reduced incoming long-wave radiation at the surface has also been reported as a potential contributor to this cooling; it results from reduced evapotranspiration and thus less water vapor in the atmosphere (Claussen et al. 2001). The cooling is however dampened by decreases in turbulent fluxes emitted from the land (decreased loss of heat and water vapor through convective processes). Those non-radiative processes are indeed well-known to often oppose the albedo-induced surface temperature changes (e.g., (Davin and de Noblet-Ducoudre 2010; Boisier et al. 2012)).

Historical land use changes have contributed to the increase in atmospheric $CO_2$ content (Section 2.4; *robust evidence*) and thus to global warming (*high agreement*). Only four modeling studies (thus *limited evidence*) have estimated this land biogeochemical effects on temperature (Pongratz et al. 2010; Brovkin et al. 2004; Matthews et al. 2004; Simmons and Matthews 2016) and the average warming reported is +0.21 ± 0.07°C; it ranges from +0.16 to +0.30°C. Recent studies however suggest that the magnitude of these simulated biogeochemical effects may be underestimated as they do not account for a number of processes such as land management, nitrogen/phosphorus cycles, land-induced changes in the emissions of $CH_4$, $N_2O$ and non-GHG emissions(Ward et al. 2014; Arneth et al. 2017; Wieder et al. 2015a; Pongratz et al. 2018). Some studies estimated radiative forcing due to both biophysical and biogeochemical effects of historical and future LULCC and found a net LULCC-induced warming (i.e., positive radiative forcing, e.g., (Ward et al. 2014; Mahowald et al. 2017). However, first the estimated biophysical radiative forcing only accounts for changes in albedo. Second the combined estimates depend on other several key modelling estimates (e.g., climate sensitivity, $CO_2$ fertilisation caused by LULCC emissions, possible synergistic effects, validity of radiative forcing concept for LULCC forcing). The comparison with the other above-mentioned modelling studies is thus difficult.

Those estimates do not account for the evolution of natural vegetation in unmanaged areas, while observations and numerical studies have reported a greening of the land in boreal regions resulting from both extended growing season and poleward migration of tree lines ((Lloyd et al. 2002; Lucht et al. 2002), Section 2.3). This greening enhances global warming via a reduction of surface albedo (winter darkening of the land through the snow-albedo feedbacks, for example (Forzieri et al. 2017)) together with enhanced photosynthesis, that is increased $CO_2$ sink, for example (Qian et al. 2010). At the same time cooling occurs due to increased evapotranspiration during the growing season. When feedbacks from the poleward migration of treeline is accounted for together with the historical land use induced biophysical effects, the biophysical annual cooling (about -0.20°C to -0.22°C on land, -0.06°C globally) is significantly dampened by the warming (about +0.13°C) resulting from the movements of natural vegetation (Strengers et al. 2010). Accounting simultaneously for both land uses and natural land cover changes in their modeling experiment reduces the cooling impacts of land uses globally in this study.

**Table 2.5 Biophysical effects on global temperature [CI 95%]. Multi-model averaged biophysical effects of large-scale land-cover changes on mean annual and global surface air temperature (°C), and their uncertainties (numbers reported are for a confidence interval of 95%). Historical and future scenarios are shown (discussed in Section 2.6.1), together with the effects of zonal deforestation in the Tropical, Temperate and Boreal latitudinal belts (discussed in Section 2.6.2.1). Historical values are based on 31 modelling studies; Future scenarios are based on 7 modelling studies; Tropical deforestation values are based on 16 modelling studies; Temperate deforestation values are based on 6 studies; and Boreal deforestation values are based on 10 modelling studies**

| | |
|---|---|
| **Tropical deforestation** | **+0.1 ± 0.3** |
| **Temperate deforestation** | **-0.4 ± 0.5** |
| **Boreal deforestation** | **-0.9 ± 0.6** |
| **Global deforestation** | **-1.2 ± 0.6** |
| **Historical LULCC** | **-0.10 ± 0.05** |
| **Future LULCC (RCP8.5)** | **-0.07 ± 0.08** |

*2.6.1.1.2 At the regional level*

The global estimates reported above mask out very contrasted regional differences. Biogeochemical effects of land-use on temperature follow the spatial patterns of greenhouse gas driven climate change with stronger warming over land than ocean, and stronger warming in northern high latitudes due to the snow and sea ice albedo feedback. Biophysical effects on the contrary are more markedly located where changes on land have occurred but are not limited to their region of occurrence. Very contrasted surface temperature changes can thus result from the combination of both effects with regions showing warming (more or less dampened by biophysical effects) while others show cooling where biophysical processes are dominant. There is *medium evidence but high agreement* that anthropogenic warming in many regions of the world has been dampened or overcome by the cooling-induced (biophysical) land use change (De Noblet-Ducoudré et al. 2012; Pongratz et al. 2010). Figure 2.22 shows that land-use induced cooling in North America and Eurasia [from -1.5°C to +0.25°C depending on the model and season] simulated at almost all seasons by seven different climate models is as large as the regional effects of anthropogenic warming experienced since pre-industrial times ((De Noblet-Ducoudré et al. 2012); comparing 1973–2002 to 1871–1900). Averaged over all agricultural areas of the world (Pongratz et al. 2010) reported a 20th century biophysical cooling of -0.10°C. (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 *very limited evidence* on the effects historical land-use changes had 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 urges for a more systematic assessment of seasonal, regional and other than mean temperature metrics in the future.

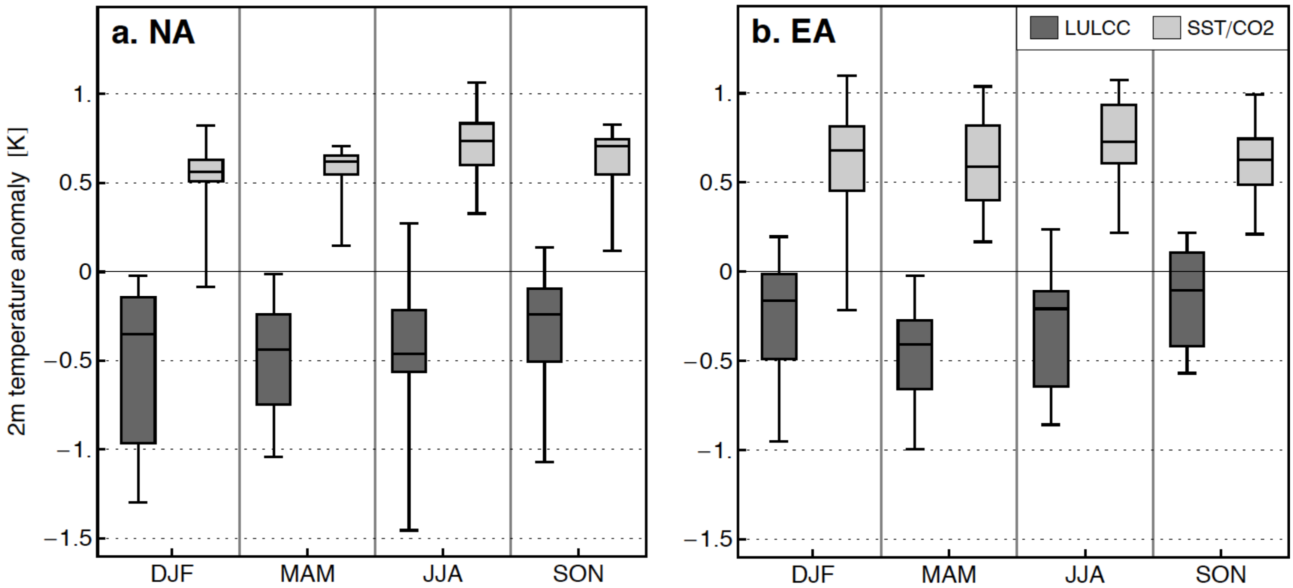


**Figure 2.22 Simulated changes in surface air temperature (K) between the pre-industrial period (1970–1900) and present-day (1972–2002) for all seasons and for a) North America and b) Eurasia (De Noblet-Ducoudré et al. 2012). Light grey boxes are the changes simulated in response to increased atmospheric greenhouse gas content between both time periods and subsequent changes in sea-surface temperature and sea-ice extent (SST/$CO_2$); dark grey boxes are the changes simulated in response to the biophysical effects of historical land use changes (LULCC). 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 (DJF). North America and Eurasia are large 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)). The figure shows that the cooling resulting from historical land use changes is almost compensating the warming resulting from climate change at all seasons and regions**

*2.6.1.1.3 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 land cover change on temperature and precipitation extremes was found to be locally as important as changes arising from increases in atmospheric $CO_2$ and sea-surface temperatures, but with a lack of model agreement on the sign of changes (Pitman et al. 2012). 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 hot 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 these three studies did not consider the effect of management changes which have been shown to be important, as discussed Section 2.6.2.

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

### 2.6.1.2 *Impacts on climate of future global land-use scenarios*

*2.6.1.2.1 At the global level*

There is *limited evidence and no agreement* on the net sign of the change in mean global annual surface temperature resulting from future RCP8.5 and RCP2.6 land use scenarios (Boysen et al. 2014; Davies-Barnard et al. 2015). The net change, estimated from five modeling studies, is -0.05 ± 0.23°C for RCP 8.5, ranging from -0.35 to +0.26°C. Such a small change is partly explained by the projected changes in areas that are smaller compared to the historical period: (Hurtt et al. 2011) indeed project that net future changes in land use will not exceed 10–30% of the historical land use change estimated between 1500 and 2005.

For those two RCP scenarios, climate models agree on a systematic biogeochemical warming resulting from the imposed land use changes, ranging from +0.1 to +0.33°C for RCP2.6 and smaller than +0.1°C for all models and RCP 8.5 (Brovkin et al. 2013).

There is *little evidence and no agreement* on the sign of the biophysical effects, under RCP8.5 scenario, that range from cooling (min: -0.20°C) for 5 modelling estimates to small warming for 2 others (max: +0.08°C) (Quesada et al. 2017a; Davies-Barnard et al. 2015; Boysen et al. 2014). Impacts of other future RCP land use scenarios have been less studied but the simulated biophysical responses are also very small (Davies-Barnard et al. 2015).

Previous IPCC scenarios (Special Report on Emission Scenarios - SRES, AR4) displayed larger land use changes than the more recent ones (RCP, AR5). There is *limited evidence* from some of those previous scenarios (SRES A2 and B1) of a small warming effect [+0.2 to +0.3°C] of land use change on mean global climate, this being dominated by the release of $CO_2$ in the atmosphere from land conversions (Sitch et al. 2005). This additional warming remains quite small when compared to the human-induced one, considering all anthropogenic influences [+1.7°C for SRES B1 and +2.7°C for SRES A2].

As for RCP8.5, land use change in the SRES A2 is dominated by a strong tropical deforestation by the end of the century. A global biophysical cooling of -0.14°C is estimated in response to this extreme change, a value that far exceeds the impacts of historical land use changes (-0.05°C) calculated using the same climate model (Davin et al. 2007). This global cooling occurs while warming was predicted over the tropical deforested regions. Oceanic feedbacks were shown to play a significant role in transmitting to other regions and latitudes the deforestation-induced cooling of the upper atmosphere. The authors derived the biophysical climatic sensitivity of their model to land use change and found an overall cooling effect of about 0.3°C $W^{-1}$ $m^{-2}$, while in response to changes in atmospheric $CO_2$ concentration their model simulates a warming of about 1°C $W^{-1}$ $m^{-2}$.

Those studies generally do not report about changes in other atmospheric variables than surface air temperature, thereby limiting our ability to assess the effects of land use changes on regional climate. Sitch et al. (2005) however report small reductions in rainfall via changes in biophysical properties of the land, following the massive tropical deforestation in SRES A2 (+ 0.5 and +0.25 mm/day respectively in the Amazon and Central Africa). They also report opposite changes (i.e., increased rainfall of about 0.25mm/day across the entire tropics and subtropics) triggered by biogeochemical effects of this same deforestation.

*2.6.1.2.2 At the regional level*

There is *no agreement* on whether future anthropogenic warming will be dampened or enhanced by the biophysical effects of land use change in many regions of the world (Boysen et al. 2014; Voldoire 2006). In regions that will undergo land use changes, dampening can be as large as -26% while enhancement is always smaller than 9% within RCP8.5 by the end of the 21st century (Boysen et al. 2014). There is even *no agreement* regarding the sign of the contribution of land use change per region (Boysen et al. 2014). Voldoire (2006) show that, by 2050 following the SRES B2 scenario the contribution of land use changes to the total temperature change can be as large as 15% in many regions northern boreal regions on both continents, and can reach about 40% in western south tropical Africa. Feddema et al. (2005) simulate large decreases in diurnal temperature range in the future (2050 and 2100 in SRES B1 and A2) following tropical deforestation in both scenarios. In the Amazon for example the diurnal temperature range is lowered by

2.5°C due to increases in minimum temperature while little change is obtained for the maximum value.

There are very few studies that go beyond analyzing the changes in mean surface air temperature. Some studies attempted to look at global changes in rainfall and found no significant influence of future land use changes (Sitch et al. 2005; Feddema et al. 2005; Brovkin et al. 2013). Quesada et al. (2017a,b) however carried out a systematic multi-model analysis of the response of a number of atmospheric, radiative and hydrologic variables (e.g., rainfall, sea level pressure, geopotential height, wind speed, soil-moisture, turbulent heat fluxes, shorwave and longwave radiation, cloudiness) to RCP8.5 land use scenario. In particular, they found a significant reduction of rainfall in 6 out of 8 monsoon regions studied (Figure 2.24) of about 1.9% to 3% (which means more than -0.5mm/day in some areas). Including land use changes in those Earth System models dampens by about 9% to 41% the projected increased rainfall with all forcings in those same regions (30% in the Global Monsoon region as defined by (Wang and Ding 2008)). In addition, they found a shortening of the monsoon season of one to four days. They conclude that the projected future increase in monsoon rains may be overestimated by those models that do not yet include biophysical effects of land use changes. Overall, the regional hydrological cycle is found to be substantially reduced and wind speed significantly strengthened in response to regional deforestation within the tropics, with magnitude comparable to projected changes with all forcings (Quesada et al. 2017b).

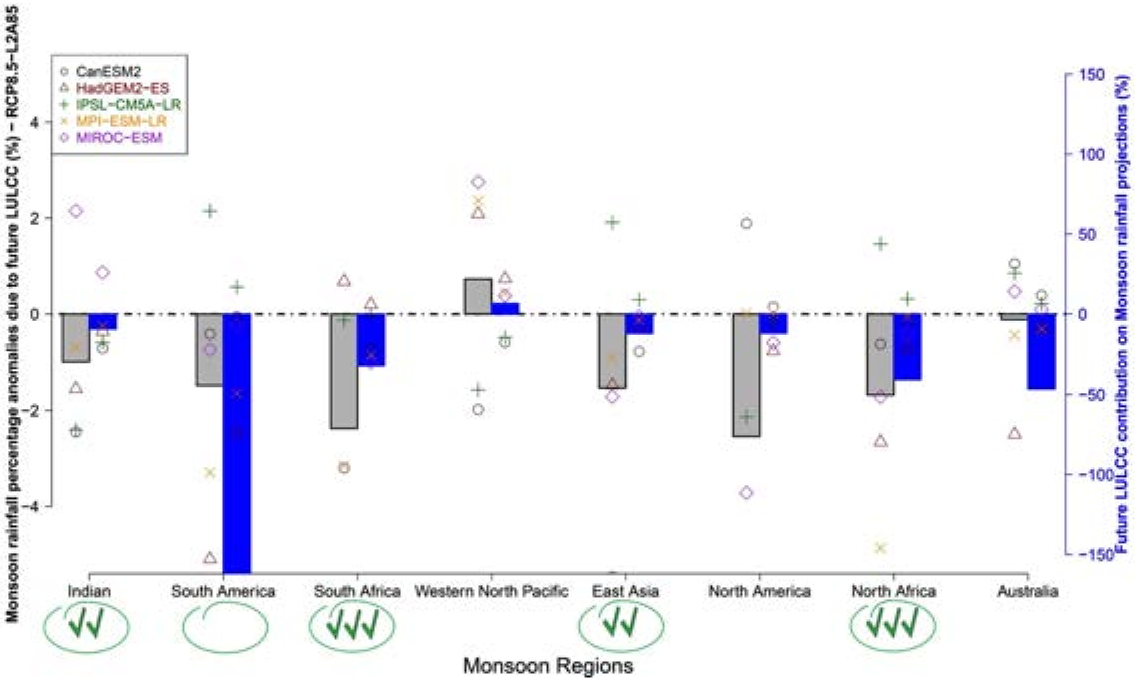


**Figure 2.23 Land use induced changes in monsoon rainfall in eight monsoonal regions (grey bars), and their relative contribution to future rainfall projections according to RCP8.5 scenario (Quesada et al. 2017b)). Differences are calculated between the end of the 21st century (2061–2100) and the end of the 20th century (1976–2005). Monsoon regions have been defined following (Yim et al. 2014). The changes have been simulated by five climate models following the LUCID-CMIP5 protocol for RCP 8.5 (Brovkin et al. 2013). Results are shown for December-January-February for southern hemisphere regions and for June-July-August for northern hemisphere regions. Grey bars (left axis) refer to the percent change in rainfall resulting from the sole land use changes; blue bars (right axis) refer to the relative contribution (in %) of land use changes to the total future change. 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% of the climate models have regional changes significant at the 66th, 75th, and 80th confidence level, respectively; green circles are added when the regional values are also significant at 90$^{th}$ confidence level**

*2.6.1.2.3 Effects on extremes*

Results from a set of climate models have shown that the impact of future LUC 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 LUC a factor of uncertainty in future climate projections.

## 2.6.2 Impacts of specific land use changes

### *2.6.2.1 Impacts of deforestation or afforestation*

Deforestation (or afforestation), wherever it occurs, triggers simultaneously warming and cooling of the surface and of the atmosphere via changes in its various characteristics (e.g., (Pitman 2003; Strengers et al. 2010; Bonan 2016)). Following deforestation, warming results from a) the release of $CO_2$ and other GHG in the atmosphere (biogeochemical impact) and subsequent increase in incoming infrared radiation at surface (greenhouse effect), b) decreased total loss of energy through turbulent fluxes (latent and sensible heat fluxes) resulting from reduced surface roughness, c) increased incoming solar radiation following reduced cloudiness that often (but not always) accompanies the decreased total evapotranspiration. Cooling occurs in response to d) increased surface albedo that reduces the amount of absorbed solar radiation, e) reduced incoming infrared radiation triggered by the decreased evapotranspiration and subsequent decrease in atmospheric water vapour. b-c-d-e are referred to as biophysical effects. Deforestation and afforestation also alter rainfall and winds (horizontal as well as vertical as will be further discussed below).

In addition, the same amount of deforestation or afforestation can trigger different climatic responses (*robust evidence; high agreement*). This is because the background climatic conditions (e.g., precipitation and snow regimes) that witness the land cover change may not be the same (Pitman et al. 2011; Hagos et al. 2014; Voldoire 2006; Feddema et al. 2005; Strandberg et al. 2018). Indeed, the magnitude and sign of local climate changes depend on a) where deforestation/afforestation occurs, b) its magnitude, c) the level of warming under which the land cover change is applied and d) the land conversion type.

Although there is not enough literature yet that rigorously compares biophysical and biogeochemical effects of deforestation or afforestation, there is evidence 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 $CO_2$, while the reverse is true when temperature is averaged at the global scale (Anderson-Teixeira et al. 2012).

The literature that discusses the effects of afforestation or reforestation on climate is more limited than for deforestation. However, most studies and observations reveal that the climatic response to afforestation has an opposite sign compared with the response to deforestation, as further discussed below. The modelling studies that combine afforestation for climate mitigation with their biophysical effects on climate are discussed in Section 2.7.

#### *2.6.2.1.1 Global and regional impacts of deforestation/afforestation in Tropical regions*

**Deforestation**

Model experiments hardly agree on the sign and magnitude of the biophysical effects of pan-tropical deforestation on the mean annual surface air temperature of the globe. The average among all modelling estimates is 0.1 ± 0.3°C (*medium evidence, no agreement*) and the range spans from -0.2°C to +1.2°C (Ganopolski et al. 2001; Snyder et al. 2004; Devaraju et al. 2015, 2018; Longobardi et al. 2016; Perugini et al. 2017).

There is however *robust evidence* and *high agreement* that such large land cover change would lead to a mean biophysical warming of the entire tropics of 0.60 ± 0.25°C (Lawrence and Vandecar 2015; Bathiany et al. 2010; Devaraju et al. 2018; Bala et al. 2007; Kendra Gotangco Castillo and Gurney 2013; Perugini et al. 2017), accompanied by an additional warming due to biogeochemical effects (release of GHG following deforestation; Section 2.4). Surface temperature response (due to biophysical effects) ranges from -0.3°C to +2.5°C depending on the models as they all have different set-up. Strict comparison between all studies (and thus assessment) is difficult as some models for example address both biogeochemical and biophysical feedbacks while some only concentrate on biophysical effects; some models include interactive oceans while others do not. The main drivers of this tropical warming are increased atmospheric $CO_2$ (when the experiment considers it) and decreased evapotranspiration resulting from the loss of trees.

There is *robust evidence* that tropical deforestation affects tropical rainfall. In case of large-scale deforestation (pan-tropical or an entire sub-continent such as the Amazon) there is *robust evidence and high agreement* that rainfall significantly decreases (Lawrence and Vandecar 2015; Perugini et al. 2017; Lejeune et al. 2015). In their meta-analysis review, (Perugini et al. 2017) reported an average simulated decrease of -288 ± 75 mm yr$^{-1}$ (95%-confidence interval). The magnitude of the deforestation-induced precipitation strongly depends on the type of land cover conversion. For instance, conversion of tropical forest to bare soil causes larger reductions in regional precipitation than conversion to pasture (respectively -470 ± 60 mm yr$^{-1}$ and -220 ± 100 mm yr$^{-1}$). Biogeochemical effects in response to pan-tropical deforestation, particularly $CO_2$ release, are generally not taken into account in those studies while they could intensify the hydrological cycle and thus precipitation (Kendra Gotangco Castillo and Gurney 2013).

Those pan-tropical changes however mask out very contrasted regional differences both in temperature and rainfall. There is *high agreement* that the impacts on regional climate resulting from the deforestation of the entire Amazon basin are larger than similar deforestation in Africa or South-East Asia (Lawrence and Vandecar 2015; Silvestrini et al. 2011; Spracklen and Garcia-Carreras 2015). Warming of the deforested areas in the Amazon is about +1.9°C, ranging from 0.1 to 3.8°C (Lawrence and Vandecar 2015) along with an Amazon drying by reducing annual mean rainfall [$-12 \pm 11\%$ according to (Spracklen and Garcia-Carreras 2015), -270 mm yr$^{-1}$ (ranging from -1204 to +394 mm yr$^{-1}$ among modeling studies) following (Lejeune et al. 2015) or –324 mm yr$^{-1}$ (ranging from -140 to -640 mm yr$^{-1}$) following (Lawrence and Vandecar 2015)]. Africa, South-East Asia and the Amazon basin differ in their sensitivity to deforestation due to regional forest intrinsic characteristics, model structure and intensity of deforestation.

Specific deforestation studies have been carried out for Africa (Hagos et al. 2014; Boone et al. 2016; Xue et al. 2016; Nogherotto et al. 2013; Hartley et al. 2016; Klein et al. 2017; Abiodun et al. 2012), southern America (Butt et al. 2011; Wu et al. 2017b; Spracklen and Garcia-Carreras 2015; Lejeune et al. 2015), South-East Asia (Ma et al. 2013b; Werth and Avissar 2005; Mabuchi et al. 2005; Tölle et al. 2017). All report decreases in evapotranspiration following deforestation (*high agreement*), resulting surface warming despite the competing effect from increased surface albedo (*high agreement*). Changes in horizontal surface winds are also reported (*high agreement*). What is mostly put forward is an increase in the land-sea thermal contrast in Africa and South America. African deforestation studies have concluded that the reduced surface friction following deforestation increases the monsoon flow, transporting water vapor further away. The increased land-ocean thermal gradient found in Southern America increases the oceanic influx; it results in a contrasted picture where moisture flux (and therefore rainfall) is reduced over the north-western part of the Amazon, while it increases over the sou part (Wu et al. 2017b).

**Afforestation**

According to the review carried out by (Perugini et al. 2017), there is *limited evidence but high agreement* that the biophysical effects of pan-tropical afforestation cool tropical surface air temperature (-0.17 ± 0.10°C, n=6 modelling estimates) and increases tropical rainfall (+41 ± 21 mm yr$^{-1}$). There is however *no agreement* (*limited evidence)* on the sign of pan-tropical deforestation biophysical impact on global surface temperature (Bathiany et al. 2010; Wang et al. 2014b). The smaller magnitude of pan tropical afforestation compare to pan tropical deforestation results from the smaller change in total forest area in the former. (Arora and Montenegro 2011) discussed the biophysical and biogeochemical cooling resulting from increasing tree cover in the tropics and conclude that tropical afforestation is three times more effective in cooling down climate than are boreal or temperate afforestation.

Large-scale afforestation scenarios of West Africa (Abiodun et al. 2012), eastern China (Ma et al. 2013a) or Saharan and Australian deserts (Ornstein et al. 2009; Kemena et al. 2018) all concluded that cooling is simulated wherever trees are grown (-2.5°C in the Sahel and -1°C in the Savannah area of West Africa, up to -8°C in western Sahara, -1.21°C over land in eastern China while cooling of the ambient air is smaller (-0.16°C). In the case of Savannah afforestation this decrease entirely compensates the GHG induced warming following the SRES A1B scenario (+1°C). West African countries thus have the potential to dampen partially, or even totally at some places, the GHG-induced warming in the deforested regions (Abiodun et al.

2012). However this is compensated by enhanced warming in adjacent countries.

Complete afforestation of the Saharan and the Australian deserts, through the large-scale deployment of irrigation and of desalination plants, came with significant increases in rainfall. Those are however not sufficient to sustain the planted forests without irrigation: in the Sahara 26–50% of the produced evapotranspiration could be recycled (Ornstein et al. 2009; Kemena et al. 2018).

*2.6.2.1.2 Global and regional impacts of deforestation/afforestation in Temperate regions*

**Deforestation**

There is *medium agreement* (based on *limited evidence)* that large-scale temperate deforestation leads to global biophysical cooling with a mean value of -0.4 ± 0.5°C ranging from -1.1°C to +0.28°C (Perugini et al. 2017; Devaraju et al. 2018, 2015; Longobardi et al. 2016; Bala et al. 2007; Snyder et al. 2004). The cooling is largely driven by the increased albedo, enhanced by the snow-albedo feedback during winter and early spring and, in some models, cooling is amplified by the response of boreal regions to temperate deforestation.

There is *no agreement* on the sign of the temperature change in temperate regions following large-scale temperate deforestation (*medium evidence*). Climate models agree on a surface biophysical cooling (*high agreement*), with a mean value of -0.7 ± 0.4°C, ranging from -0.1 to -1.3°C (Devaraju et al. 2015; West et al. 2011; Snyder et al. 2004; Gattuso et al. 2018; Dümenil Gates and Ließ 2001). Those modeling results however are not supported by observations as those suggest that the resulting biophysical effect is net mean warming of the temperate latitudes(Alkama and Cescatti 2016; Lee et al. 2011; Duveiller et al. 2018a) although (Alkama and Cescatti 2016) report cooling in many Eurasian regions while systematic warming is found in North America. The existence of both cooling and warming is due to the fact biophysical effects in the temperate regions compensate one another: the albedo-induced cooling, enhanced by winter snow, compensates the warming resulting from decreased evapotranspiration and roughness (Davin and de Noblet-Ducoudre 2010; Li et al. 2016). Disagreement between models and observations may result from a large sensitivity of climate models to albedo changes, and/or to the scale of the deforestation that is pan-temperate in the models and multi-local in the observations.

The lack of agreement at the annual scale among the climate models is however masking *rising agreement* (the *evidence is yet still limited*) regarding seasonal and diurnal impacts of deforestation at those latitudes. There is *high agreement* that temperate deforestation leads to summer warming in response to decreased loss of energy via turbulent fluxes (more specifically evapotranspiration), while albedo-induced cooling is found during winter time. In addition there is growing evidence from observations that daytime temperatures are warmer following deforestation while night time temperatures are cooler (Alkama and Cescatti 2016; Duveiller et al. 2018b; Lee et al. 2011; Peng et al. 2014). More recently (Lejeune et al. 2018) found systematic warming of the hottest summer days following historical deforestation in the northern mid latitudes that needs to be further examined.

**Afforestation**

Perugini et al. (2017) reported a mean annual cooling of the temperate latitudinal band following afforestation when values are based on observations(Alkama and Cescatti 2016; Lee et al. 2011; Duveiller et al. 2018a) while warming is reported from the sole modeling study they found that looked at pan-temperate afforestation. Those results are consistent with (and opposite to) the effects of deforestation discussed above.

There is *robust evidence and high agreement* that afforestation in North America or in Europe cools surface climate during summer time, especially in regions that remain wet enough to allow trees to transpire (Bright et al. 2017; Zhao and Jackson 2014; Gálos et al. 2011, 2013; Wickham et al. 2013; Ahlswede and Thomas 2017; Anderson-Teixeira et al. 2012; Anderson et al. 2011; Chen et al. 2012; Strandberg et al. 2018). Cooling results from a significant decrease in the loss of energy through convective heat fluxes (sensible and latent heat; (Anav et al. 2010) although most studies only report on their simulated large increases in evapotranspiration. In temperate regions that suffer from water deficit the change in evapotranspiration following afforestation will be insignificant while the decreased surface albedo will favor surface warming. Some studies also report cooling of spring and fall, or mean annual cooling. Others report winter warming due to the decreased surface albedo, which is more pronounced during the snow season. (Chen et al. 2012) and (Gálos et al. 2011, 2013) found increased precipitation in cooler afforested regions respectively in the USA and in Europe. (Strandberg et al. 2018) found that the August 2003 and July 2010 heat-waves could have been largely mitigated if Europe had been largely afforestated.

Combining large-scale afforestation of western Europe and climate change scenario (SRES A2) (Gálos et al. 2013) found a relatively small damping potential of additional forest on ambient air temperature (land-induced cooling of about -0.5°C versus GHG-induced warming larger than 2.5°C). Influence on rainfall was however much larger and significant. While Germany, France and Ukraine experienced decreases in annual rainfall following warming, afforestation can revert this signal in Germany and significantly dampen it in both France & Ukraine. In addition, the warming-induced increase in the number of dry days is also dampened by afforestation while the number of extreme precipitation events is amplified.

Less extreme afforestation has been tested in the warmer world predicted by RCP 8.5 scenario, and the net impact of afforestation was examined, combining both biophysical and biogeochemical effects (Sonntag et al. 2016; Sonntag 2018). 8 M $km^2$ of forests were added globally, following the land use RCP 4.5 scenario. The global cooling resulting from this increase in forest area is too small (-0.27°C annually) to dampen the RCP 8.5 warming. It however reaches about -1°C in some temperate regions and -2.5°C in boreal ones. This is accompanied by a reduction in the number of extremely warm days. Detailed regional analysis of this global run however shows that dampening in the densely populated areas is quite small, while it can be very large in sparsely populated areas.

*2.6.2.1.3 Global and regional impacts of deforestation/afforestation in Boreal regions*

**Deforestation**

There is *medium evidence* and *high agreement* that large-scale boreal deforestation leads to global biophysical cooling with a mean value of -0.9 ± 0.5°K ranging from -2.8°K to -0.23°K (Claussen et al. 2001; Snyder et al. 2004; Devaraju et al. 2015, 2018; Longobardi et al. 2016; Bala et al. 2007; Dass et al. 2013). This cooling is essentially driven by the increased albedo, enhanced by the snow-albedo feedback during winter and early spring and, in some models.

In addition boreal regions are also experiencing a biophysical cooling of -*2.2* ± 0.*8*°K (Bathiany et al. 2010; Devaraju et al. 2015; Dass et al. 2013; Ganopolski et al. 2001; Snyder et al. 2004; West et al. 2011).

**Afforestation**

Future global afforestation has been tested for SRES A2 in a fully coupled global climate model. Afforestation of either 50% or 100% of the total agricultural area has been gradually prescribed between years 2011 and 2060. In addition boreal, temperate and tropical afforestation have been tested separately. Both biophysical and biogeochemical effects have been accounted for (Arora and Montenegro 2011).The net impacts of afforestation was quite marginal compared to the GHG-induced global warming (+3°C at the end of the 21st century) but was indeed the expected cooling effect (from -0.04°C to -0.45°C depending on the location and magnitude of the additional forest cover). Consistent with previous experiments, increasing forests in boreal regions induced biophysical warming and biogeochemical cooling while increasing tree cover in the tropics led to both biophysical and biogeochemical cooling. The authors conclude that tropical afforestation is three times more effective in cooling down climate than are boreal or temperate afforestation.

***2.6.2.2 Impacts of changes in agriculture 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. Those may affect water and energy fluxes to the atmosphere, and thus temperature and rainfall, to the same extent as changes in land cover do as discussed in (Luyssaert et al. 2014) and (Wilfert et al. 2016).

We report herein on the effects of one specific land management (irrigation) that has been substantially studied, and one attempt to manage solar radiation via increases in crop albedo (geoengineering the land).

There are a number of other practices that exist and matter for climate, but there is not enough literature available for an assessment as only one to few papers are generally found per practice (e.g., (Jeong et al. 2014b) for double cropping, (Bagley et al. 2017) for the positioning in time of the seasonal cycle).

*2.6.2.2.1 Irrigation*

There is substantial literature on the effects of irrigation on local, regional and global climate as this is a major land management. There is *robust evidence and high agreement* that irrigation increases total evapotranspiration, increases the total amount of water vapor in the atmosphere, and decreases mean surface daytime temperature within the irrigated area and during the time of irrigation (Bonfils and Lobell 2007; Alter et al. 2015; Chen and Jeong 2018; Christy et al. 2006; Im et al. 2014; Im and Eltahir 2014; Mueller et al. 2016). Decreases in maximum daytime temperature can locally be as large as -3°C to -8°C (Alter et al. 2015; Cook et al. 2015a; Han and Yang 2013; Huber et al. 2014; Im et al. 2014). Estimates of the contribution of irrigation to past historical trends in ambient air temperature vary between -0.07°C and -0.014°C/decade in Northern China (Han and Yang 2013; Chen and Jeong 2018) while being quite larger in California (-0.14°C to -0.25°C per decade; Bonfils and Lobell, 2007). Surface cooling results from increased energy being uptaken from the land via larger evapotranspiration rates, while decreases in ambient air temperature is driven by decreased sensible heat flux (less heat is brought to the atmosphere through convection). In addition there is evidence that the irrigation-induced evaporative cooling can locally mitigate the effect of heatwaves (Wim et al. 2017; Mueller et al. 2015).

There is *no agreement* on changes in nighttime temperatures as discussed in (Chen and Jeong 2018) who summarised the findings from observations in many regions of the World (India, China, North America and eastern Africa; Figure 2.24). Where nighttime warming is found (Chen and Jeong 2018; Christy et al. 2006), two explanations are put forward: the first is an increase in incoming long-wave radiation in response to increased atmospheric water vapour content (greenhouse effect); the second is an increased storage of heat in the soil during daytime, because of the larger heat capacity of a moister soil, heat that is then released to the atmosphere at night.

There is *robust evidence* that allowing irrigation enhances rainfall although there is *no agreement* on where this increase occurs. When irrigation occurs in Sahelian Africa, during the monsoon period, rainfall is decreased over the irrigated areas (*high agreement*) and increases south-west if the crops are located in western Africa (Alter et al. 2015) and east / north-east when crops are located further East in Sudan (Im et al. 2014; Im and Eltahir 2014). Irrigation-induced cooler surfaces in the Sahel inhibits convection and creates an anomalous descending motion over crops that suppresses rainfall but influences the circulation of monsoon winds. Irrigation in India occurs prior to the start of the monsoon season and the resulting land cooling decreases the land-sea temperature contrast. This can delay the onset of the Indian monsoon and decrease its intensity (Niyogi et al. 2010; Guimberteau et al. 2012). Results from De Vrese et al. (2016) modeling study suggests that part of the excess rainfall triggered by Indian irrigation falls eastward, in the horn of Africa.

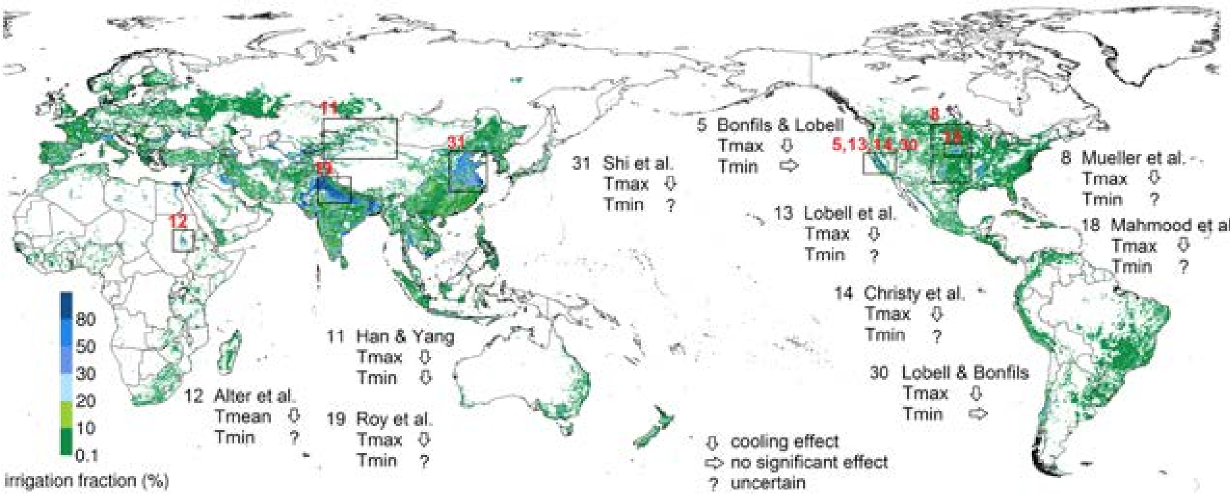


**Figure 2.24 Irrigation impacts on the sign of the surface temperature change (mean, maximum or minimum values) from observational studies as reviewed in (Chen and Jeong 2018). Colors refer to the areas equipped for irrigation around the year 2005, expressed as a percentage of total area (irrigation fraction; (Siebert et al. 2013)). The boxes indicate the research regions of the observational studies. References are (Alter et al. 2015; Han and Yang 2013; Roy et al. 2007; Shi et al. 2013; Bonfils and Lobell 2007; Lobell et al. 2008; Lobell and Bonfils 2008; Christy et al. 2006; Mueller et al. 2015; Mahmood et al. 2006)**

*2.6.2.2.2 Albedo changes*

There is *medium evidence and high agreement* that increasing surface albedo in cropland areas will locally reduce surface temperatures (Carrer et al. 2018; Crook et al. 2015; Hirsch et al. 2017; Lobell et al. 2006; Millar et al. 2018; Ridgwell et al. 2009; Singarayer et al. 2009; Singarayer and Davies-Barnard 2012; Sousa et al. 2018; Wilhelm et al. 2015; Seneviratne et al. 2018). Three methods have been proposed and are summarised in (Seneviratne et al. 2018): choosing specifically 'bright' varieties, abandoning tillage (as in (Lobell et al. 2006; Davin et al. 2014)) or using greenhouses (as in (Campra et al. 2008)). In addition and depending on the location, Carrer et al. (2018) have suggested that intercropping may also have a brightening effect in Europe where soils are darker than any cultivated plant.

Whatever the solution chosen, the induced reduction in absorbed solar radiation cools the land, more specifically during the hottest summer days (Davin et al. 2014; Wilhelm et al. 2015) (Figure 2.25). Changes in temperature however are essentially local and seasonal (limited to crop growth season) or sub-seasonal (when resulting from intercropping or tillage suppression). This solar radiation management action thus cannot be one of the mitigation solutions but holds the potential to help couteracting warming in cultivated areas during summer time.

The few studies that have gone beyond looking at just temperature have also reported significant changes in rainfall, specially within the indian and asian monsoon regions. They however claim that the albedo-induced cooling benefits can be overcome by decreases in rainfall that are harmful for crop productivity (Seneviratne et al. 2018).

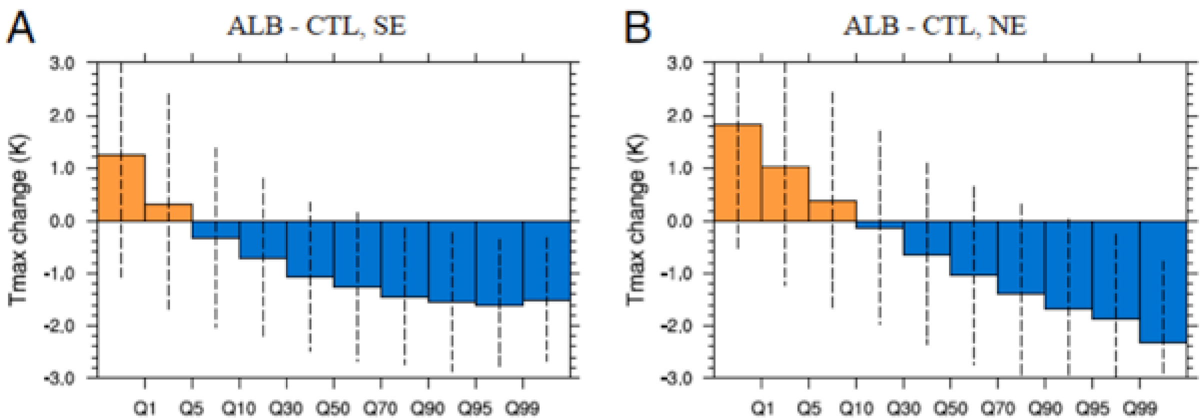


**Figure 2.25 Change in daily maximum temperature (K) resulting from increased surface albedo in Europe, following ploughing suppression (Davin et al. 2014). Changes are simulated for different quantiles of the daily maximum temperature distribution. Differences are calculated at each grid point with more than 60% of cropland and for each summer day (within July–August) over the period 1986–2009. Differences are then averaged for each quantile of daily maximum temperature defined based on the control experiment (without ploughing suppression – CTL. The experiment with enhanced albedo is referred to as ALB). 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)**

#### *2.6.2.3 Urbanisation-induced climate and weather changes*

The presence of cities alters the local and neighbouring atmospheric conditions. The most well-known phenomena is the urban heat island (UHI). It results from urban air temperatures being substantially larger than that in the surrounding rural areas. This excess heat in cities is due to three main factors: 3-D urban geometry that trap radiation, thermal characteristics of impervious surfaces and the release of anthropogenic heat (Bohnenstengel et al. 2014; Ichinose et al. 1999; Ma et al. 2017a). The magnitude and diurnal amplitude of the UHI varies from one city to another as it depends on the local background climate (Wienert and Kuttler 2005; Zhao et al. 2014; Ward et al. 2016). Other phenomenon have been reported, such as the urban dryness island, referring to lower relative humidity in cities relative to more rural locations where evapotranspiration is larger as more vegetation is present (Bader et al. 2018); or the urban wind island as cities have been shown to experience slower wind speeds compared to their adjacent suburban and countryside due to their larger roughness length (Bader et al. 2018; Wu et al. 2017a).

There is *limited evidence but high agreement* that the global annual mean surface air temperature response to urbanisation is negligible ([<0.01 – 0.06 °C], (Zhang et al. 2013a; Chen et al. 2016b; Hansen et al. 2010; Parker 2006)) owing to their tiny relative fractional coverage (< 1 %, (Wang et al. 2017a; Schneider et al. 2010)) of the world's surface. Recent studies however suggest that the urban anthropogenic heat has been responsible of a warming by as much as 1°C in mid and high latitude in winter and autumn over North America and Eurasia, and of an equatorward shift of the winter mid-latitude jet, with increasing westerly wind at 20°N and decreasing westerly wind at 40°N (*limited evidence*; (Lamptey 2009; Chen et al. 2016a)).

There is *robust evidence and high agreement* that urbanisation increases mean annual temperature where cities grow and in their surroundings (Figure 2.26), ([0.19°C – 2.60 °C],(Torres-Valcárcel et al. 2015; Li et al. 2018a; Doan et al. 2016; Kim et al. 2016; Sun et al. 2016; Bader et al. 2018; Chen et al. 2016a; Alizadeh-Choobari et al. 2016; Elagib 2011; Fujibe 2009; Liao et al. 2017; Founda et al. 2015; Rafael et al. 2017; Lokoshchenko 2017; Hinkel and Nelson 2007; Chrysanthou et al. 2014)). The annual-mean maximum temperature is substantially less affected by urbanisation than the minimum temperature (*robust evidence, high agreement*; (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. 2017b; Hamdi 2010; Arsiso et al. 2018; Elagib 2011; Lokoshchenko 2017; Robaa 2013). There is *robust evidence and high agreement* of synergistic interactions between the UHI and heat wave episodes making the heat wave more intense [1.22°C–4 °C ] in urban than rural areas and the nocturnal UHI during heat wave stronger than its climatological mean (Li and Bou-Zeid 2013; Wang et al. 2017b; Hamdi et al. 2016; Founda and Santamouris 2017).

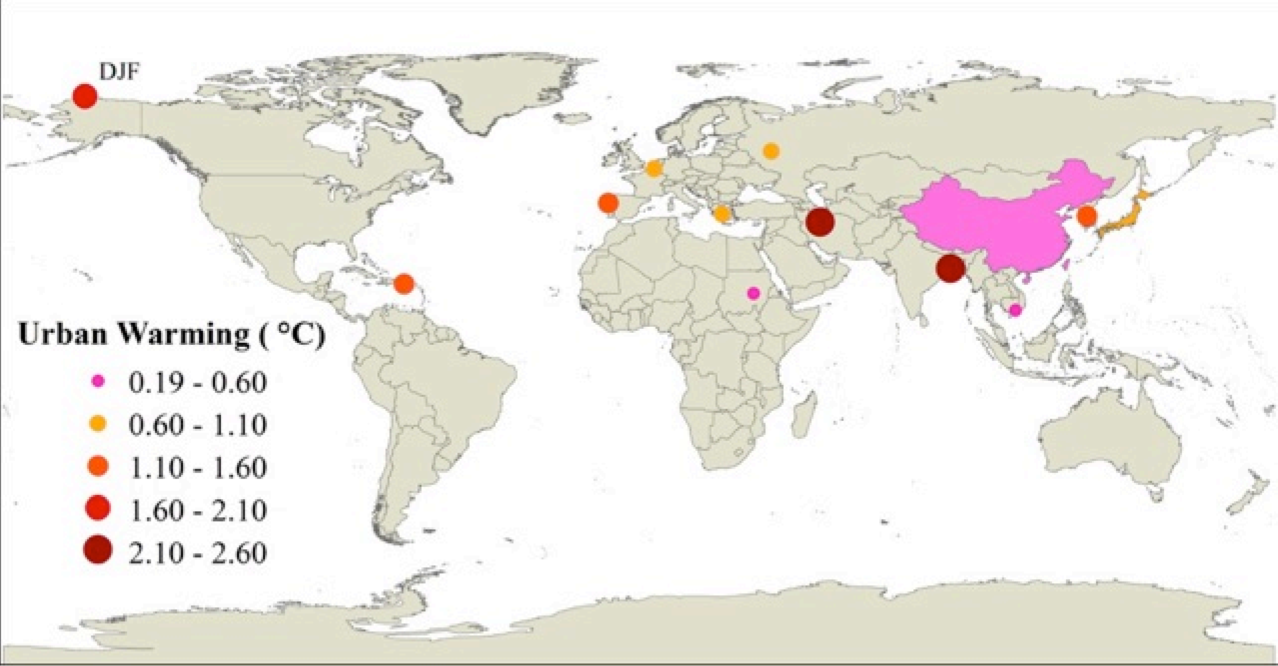


**Figure 2.26 Change in annual mean surface air temperature resulting from urbanisation (°C). Colour and size of the circles refer to the magnitude of the change. This map has been compiled using the following studies: (Torres-Valcárcel et al. 2015; Li et al. 2018a; Doan et al. 2016; Kim et al. 2016; Sun et al. 2016; Alizadeh-Choobari et al. 2016; Elagib 2011; Fujibe 2009; Founda et al. 2015; Rafael et al. 2017; Lokoshchenko 2017; Hinkel and Nelson 2007; Bader et al. 2018)**

Theoretical analysis (Han et al. 2011) and regional climate models using urban canopy parameterisations (Ganeshan and Murtugudde 2015; Li et al. 2017a; Pathirana et al. 2014; Song et al. 2016; Trusilova et al. 2008; Zhong and Yang 2015; Zhu et al. 2017b; Seino et al. 2018; Zhong et al. 2017; Ooi et al. 2017) have explained how urbanisation influences the precipitation patterns near urban centres. There is *medium evidence and medium agreement* that urban areas induce increases in mean but also in extreme precipitations over and downwind of the city in different regions of the world, especially in the afternoon and early evening when convective activity is the largest: Atlanta(Haberlie et al. 2014; McLeod et al. 2017); different inland and coastal US cities (Ganeshan and Murtugudde 2015); Dutch coastal cities (Daniels et al. 2016); Hamburg (Schlünzen et al. 2010); Shanghai (Liang and Ding 2017), Beijing (Dou et al. 2015). Accounting for increased aerosols concentrations in urban areas however can interrupt precipitation formation process and thereby reduce heavy rainfall (Zhong et al. 2017; Daniels et al. 2016). Urban areas also affect other components of the water cycle, increasing the evapotranspiration demand for plants in cities by as much as 10% (Zipper et al. 2017)and increasing the surface runoff (Hamdi et al. 2011).

There is *no agreement* on how the UHI will evolve under climate change conditions because several studies using different methods (dynamical/statistical downscaling, GCM/RCM, urban RCM/offline impact models) report contrasting results with both increasing and decreasing UHI (McCarthy et al. 2010; Oleson et al. 2011; Oleson 2012; Adachi et al. 2012; Hamdi et al. 2014; KUSAKA et al. 2012b,a; Mccarthy et al. 2012; Arsiso et al. 2018; Früh et al. 2011; Hatchett et al. 2016; Sachindra et al. 2016; Rafael et al. 2017; Lauwaet et al. 2015; Hoffmann et al. 2015; Lemonsu et al. 2013; Wilby 2003). There is *robust evidence and high agreement* however that the impact of future urbanisation scenarios either under present climate or combined with future climate conditions will increase warming in different climatic regions (Mahmood et al. 2014; Kaplan et al. 2017; Doan et al. 2016; Li et al. 2018a; Georgescu et al. 2013; Argüeso et al. 2014; Kim et al. 2016; Kusaka et al. 2016; Grossman-Clarke et al. 2017) with a strong impact on minimum temperatures that could be comparable to the climate change signal only for the near future over western Europe (+0.6 °C, (Berckmans et al. 2018)).

### 2.6.3 Amplifying / dampening climate changes via land responses

Section 2.2 illustrates the various mechanisms through which land can affect the atmosphere and thereby climate and weather. Section 2.3 illustrates the many impacts climate changes have on the functioning of land ecosystems. Sections 2.6.1 to 2.6.2 show the various effects changes in land cover (e.g., afforestation/deforestation), or uses (e.g., irrigation) have on atmospheric states (e.g., air temperature, rainfall) or processes (e.g., convection, monsoon). Land has thus the potential to dampen or amplify either the greenhouse-induced global climate warming or its regional consequences (e.g., drought or moistening, warming or cooling), as schematically illustrated Figure 2.27. It can be used as a tool to mitigate (attenuate) some regional climatic consequences of global warming (e.g., extreme weather events), and not only via increasing the capacity of land to absorb $CO_2$.

Such feedbacks can only be consistently assessed with climate models that account for dynamic vegetation, that is for the responses to climate changes of a) plants' growth and production, and b) the geographical distribution of vegetation. Those processes are not accounted for in all climate projections (and almost absent from regional climate ones). In addition, the specific demonstration of how those processes feedback on the initial greenhouse-induced warming is not often studied, thus limiting our ability here to carry out the assessment.

Section 2.4 discusses the effects future climate change scenarios have on the evolution of the capacity of the land to absorb part of the anthropogenic $CO_2$, which then controls the sign of the feedback on initial global warming (amplification (resp. dampening) if this capacity is reduced (resp. increased)).

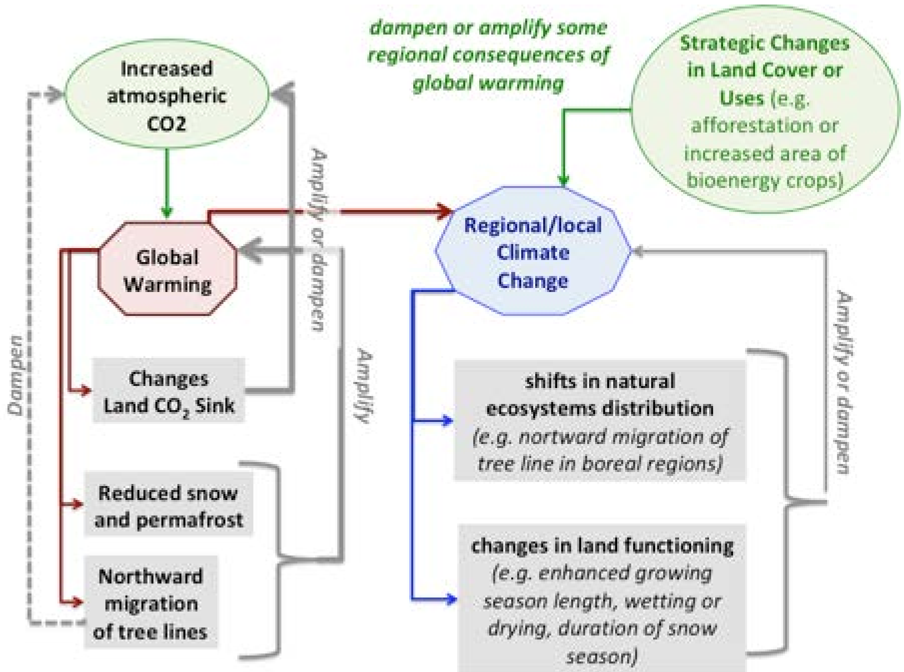


**Figure 2.27 Schematics of the various ways Land has been shown, in the literature, to either amplify or dampen the initial climate change, at the global scale (left panel) or at the regional/local levels (right panel). Green arrows and boxes refer to what we consider herein as imposed changes. Grey 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**

***2.6.3.1 Effects of climate-induced changes in land productivity and shifts in vegetation distribution***

There is *robust evidence and high agreement* that, 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 2.3) will have positive feedbacks both on global and regional annual warming (Garnaud and Sushama 2015; Jeong et al. 2014a; O'ishi and Abe-Ouchi 2009; Port et al. 2012; Strengers et al. 2010) through biophysical feedbacks. This warming response to increased greening is supported by recent analysis of satellite observations (Forzieri et al. 2017) and has also been put forward in paleoclimate studies of e.g. the last glacial inception (De Noblet et al. 1996; Kageyama et al. 2004).

As an example, (Jeong et al. 2014a; Garnaud and Sushama 2015) a significant reduction in surface albedo especially in late winter and early spring in response to greening (Figure 2.28). The presence of more trees or increased biomass at those latitudes darkens the snowy ground and allows more solar radiation to be absorbed (Loranty et al. 2014). As a consequence, the duration and the amount of snow on the ground reduce, thereby enhancing warming during winter and spring. On the contrary during late spring and early summer denser vegetation increases the loss of energy from the surfaces in the form of evapotranspiration (or latent heat flux) and this tends to dampen the GHG-induced warming regionally. There is yet *limited evidence but growing agreement* that greening-induced warming in boreal regions is constrained to some seasons. Limited evidence results from the absence of discussion regarding seasonal variables in the vast majority of published papers.

At the annual as at the global scale the snow-albedo-induced warming remains the dominant signal in all modeling studies. (Snyder and Liess 2014) however have warned the community that the intensity of this feedback may be overestimated as is the growth of vegetation in boreal regions in climate models.

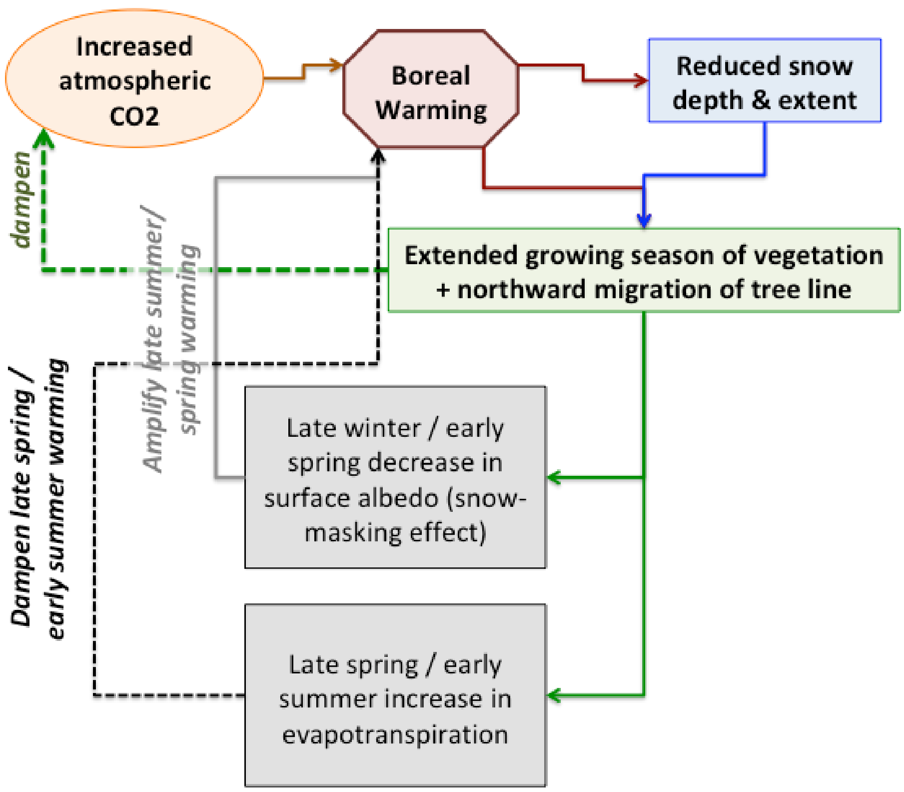


**Figure 2.28 Schematic illustration of how climate-induced changes in the boreal regions on boreal warming. The sign of the feedbacks depends on the season, although annually global warming is further enhanced in those regions. Dashed lines illustrate negative feedbacks while plain lines indicate positive feedbacks**

In the tropics there is *medium evidence but high agreement* that, in regions where global warming provokes reduction (resp. increase) in rainfall, the induced decrease (resp. increase) in biomass production leads to enhanced (resp. reduced) warming (Port et al. 2012; Wu et al. 2016; Yu et al. 2016).

As an example, (Port et al. 2012) simulated decreases in tree cover and shortened growing season in the Amazon, despite the $CO_2$ fertilisation effects, in response to both future tropical warming and reduced precipitation (Figure 2.29). This browning of the land decreases both evapotranspiration and atmospheric humidity, with opposite effects on surface warming. Evapotranspiration-induced warming in enhanced via decreases in cloudiness that increases incoming solar radiation.

There is *no agreement* on feedbacks on rainfall in the tropics, over the areas of vegetation changes, as greening-induced increases or decreases in precipitation depend on where the greening occurs and on the season (as discussed in Section 2.6.2). There is however *robust evidence and high agreement* that greenhouse-induced vegetation growth in the southern Sahel increases African monsoon rains (Port et al. 2012; Wu et al. 2016; Yu et al. 2016). Confidence on the right 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).

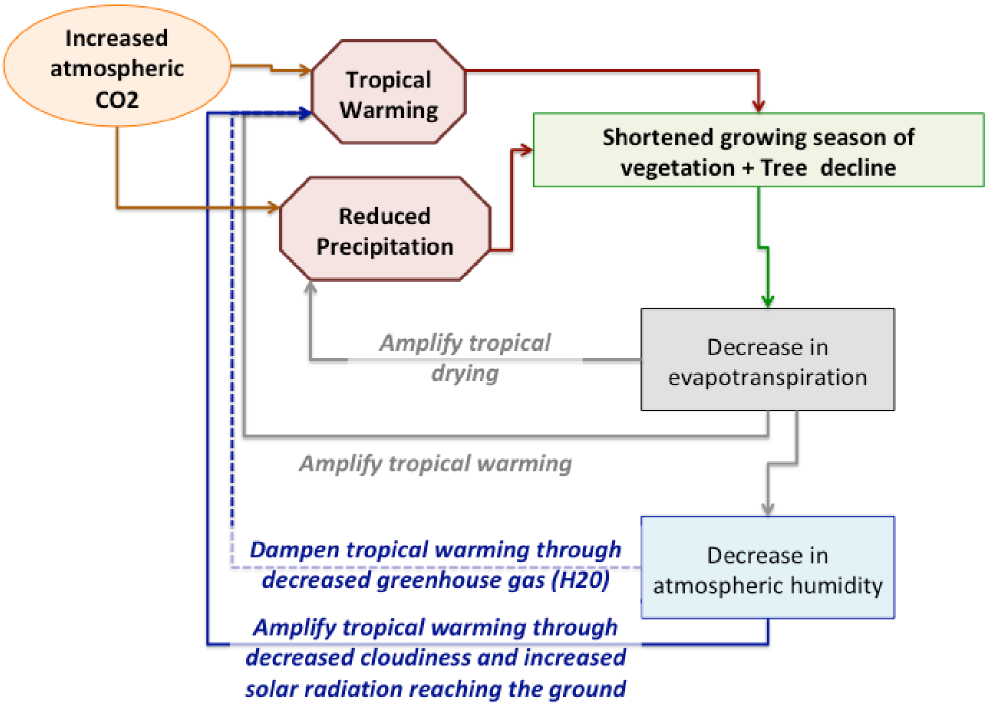


**Figure 2.29 Schematic illustration of how climate-induced changes in vegetation distribution in the Amazon region feedback on tropical warming and drying in the modeling experiment of (Port et al. 2012). Many feedbacks occur simultaneously and in opposite direction. Dashed (resp. plain) lines are for negative (resp. positive) feedbacks**

#### *2.6.3.2 Feedbacks related to high-latitude climate-induced land-surface changes*

In high latitudes, snow albedo and permafrost carbon feedbacks are not only the most well-known surface-related climate feedbacks, but also the most important ones because of their large-scale impacts.

There is *robust evidence and high agreement* that, 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 (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. (Flanner et al. 2011) and (Singh et al. 2015) indeed evaluated the all-sky global land snow shortwave radiative effect to be around -2.5±0.5 W m$^{-2}$. In the Southern Hemisphere, the Antarctic contribution (≈ -3.1±0.3 W m$^{-2}$) is by far dominant, while in the Northern Hemisphere, this is essentially attributable to seasonal snow (≈ -1.5±0.5 W m$^{-2}$), with a smaller contribution (≈ -0.45±0.10 W m$^{-2}$) from glaciated areas. Another reason is the sensitivity of snow cover to temperature. (Mudryk et al. 2017) recently showed that in the high latitudes, climate models tend to correctly represent this sensitivity, while in mid-latitude and alpine regions, the simulated snow cover sensitivity to temperature variations tends to be biased low. In total, the global snow albedo feedback is about 0.1 W $m^{-2}$ $K^{-1}$, which amounts to about 7% of the strength of the globally dominant water vapor feedback (e.g., (Thackeray and Fletcher 2015)). While climate models do represent this feedback, a persistent spread in the modeled feedback strength has been noticed (Qu and Hall 2014) and, on average, the simulated snow albedo feedback strength tends to be somewhat weaker than in reality (Flanner et al. 2011; Thackeray and Fletcher 2015) (*moderate confidence*). Various reasons for the spread and biases of the simulated snow albedo feedback have been identified, notably inadequate representations of vegetation masking of snow in forested areas (Loranty et al. 2014; Wang et al. 2016c; Thackeray and Fletcher 2015).

There is *robust evidence and high agreement* that, following permafrost decay in a warmer climate, the resulting emissions of carbon dioxide and/or methane (caused by the decomposition of organic matter in previously frozen soil) will produce additional greenhouse-gas induced warming. There is however substantial uncertainty on the magnitude of this feedback, although recent years have seen large progress in its quantification. Lack of agreement results from several critical factors among which uncertainties regarding a) the size of the permafrost carbon pool, b) its decomposability, c) the magnitude, timing and pathway of future high-latitude climate change and d) the correct identification and model representation of the processes at play (Schuur et al. 2015). The most recent comprehensive estimates establish a total soil organic carbon storage of about 1500 ± 200 PgC (Hugelius et al. 2014, 2013; Olefeldt et al. 2016), which is about 300 PgC lower than previous estimates (*limited evidence*). Important progress has been made in recent years at incorporating permafrost-related processes in complex Earth System Models (e.g., (McGuire et al. 2018), but representations of some critical processes such as thermokarst formation are still in their infancy (Schuur et al. 2015). Recent model-based estimates of future permafrost carbon release, based on offline land simulations for example (Koven et al. 2015; McGuire et al. 2018) have converged on the insight that substantial carbon release of the coupled vegetation-permafrost system will probably not occur before about 2100, because carbon uptake by increased vegetation growth will initially compensate for that released from permafrost (*limited evidence but high agreemen*t).

#### *2.6.3.3 Feedbacks related to climate-induced changes in soil moisture*

There is *medium evidence and high agreement* that soil moisture conditions influence the frequency and magnitude of extremes. Observational evidence indicates that dry soil moisture conditions favor heat-waves, in particular in regions where evapotranspiration is limited by moisture availability (Mueller and Seneviratne 2012; Quesada et al. 2012; Miralles et al. 2018; Geirinhas et al. 2018; Miralles et al. 2014; Chiang et al. 2018).

In future climate projections, it has been suggested that soil moisture plays an important role in the projected amplification of extreme heat-waves and drought in many regions of the world (*medium evidence*, *medium agreemen*t; (Seneviratne et al. 2013; Vogel et al. 2017; Miralles et al. 2018)). Quantitative estimates of the actual role of soil moisture feedbacks are however very uncertain due to the *low confidence* in projected soil moisture changes (IPCC 2013, WGI SPM), weaknesses in the representation of soil moisture-atmosphere interactions in climate models (Sippel et al. 2017; Miralles et al. 2018) and methodological uncertainties associated with the soil moisture prescription framework commonly used to disentangle the effect of soil moisture on changes in temperature extremes (Hauser et al. 2017).

There is *limited evidence that,* where soil moisture decreases in response to climate change in the subtropics and temperate latitudes, this drying is enhanced by the existence of soil moisture feedbacks (Berg et al. 2016). The initial warming-induced decrease in precipitation and increase in potential evapotranspiration and latent heat flux leads to decreased soil moisture at those latitudes and can potentially amplify both. Such feature is consistent with evidence that in a warmer climate land and atmosphere will be more strongly coupled via both the water and the energy cycles (Dirmeyer et al. 2014; Guo et al. 2006). This increased sensitivity of atmospheric response to land perturbations implies that changes in land uses and cover are expected, in the future, to have more impact on climate than they do today.

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

### 2.6.4 Land-induced teleconnections, non-local and down-wind effects

There is *robust evidence but medium agreement* that land use changes, wherever they occur, generate perturbations away from the area of change. 'Away' can be as close as neighbouring regions for example (Ma et al. 2013b; McLeod et al. 2017; Abiodun et al. 2012), or as far as areas in other latitudinal bands or very contrasted climate zones (e.g., (Lean and Warrilow 1989; Cowling et al. 2009; Davin and de Noblet-Ducoudre 2010; Devaraju et al. 2015; Lawrence and Vandecar 2015; Laguë and Swann 2016; De Vrese et al. 2016; Devaraju et al. 2018). Evidence of existing teleconnection is impossible to get from observations as it would mean being able to isolate a single perturbation and its specific atmospheric response worldwide, means through which teleconnections occur are three-fold.

(1) The most well known one is via changes in atmospheric $CO_2$ content as this greenhouse gas is well mixed in the atmosphere and has effects on the downwelling long-wave radiation (see Section 2.4).

(2) Any land cover change impacts surface temperature, surface roughness and moisture and thus the magnitude of convective fluxes that take energy out of the land, and the partitioning between those fluxes (i.e., the respective magnitude of latent and sensible heat fluxes; see Section 2.2). Those changes impact the thermal, moisture and pressure gradients (contrasts) between the area of change and its neighbours, and thereby the amount of heat, water vapor and pollutants flowing in or out of the area. This generates up-wind or down-wind changes e.g. (Ma et al. 2013b; McLeod et al. 2017; Abiodun et al. 2012).

(3) Land-induced changes are not limited to the ambient air (the lower part of the boundary layer) but can reach the upper atmosphere via changes in radiation and convection. Such changes trigger perturbations in large-scale atmospheric transport, such as the transport of atmospheric energy and water towards the pole or across the equator (Cowling et al. 2009; Laguë and Swann 2016; Feddema et al. 2005). Land-induced temperature changes have also been reported over oceans in response to large-scale vegetation changes (Cowling et al. 2009; Davin and de Noblet-Ducoudre 2010; Wang et al. 2014b).

For example (Devaraju et al. 2018), using two different climate models and similar latitudinal deforestation scenarios, have demonstrated that boreal (resp. temperate) deforestation significantly affect ambient air temperature in the tropics and in the temperate (resp. boreal and tropical) regions. Temperate cooling in both models results from boreal deforestation and varies between -0.8°C and -1.5°C. The response of boreal regions to temperate deforestation varies from warming of about +0.3°C to cooling of -1°C depending on the model. (Devaraju et al. 2015) have also reported shifts in the latitudinal position of the Intertropical Convergence Zone (ITCZ) in response to boreal and temperate deforestation. Those have significant consequences on rainfall in many monsoon areas of the world.

Lawrence and Vandecar (2015) have reviewed a number of tropical deforestation modeling experiments and reported that tropical deforestation, whether regional (in Africa, the Amazon, or Asia) or latitudinal (the entire tropics being deforested), has impacts in many subtropical and temperate regions as illustrated in Figure 2.30 (Cowling et al. 2009; Laguë and Swann, 2016) have also reported changes in the transport of atmospheric energy towards the pole or across the equator in response to large-scale greening of the Earth. In (Cowling et al. 2009) an additional warming of about 3°C in the 75°N –90°N latitudinal band was attributed to increased atmospheric inflow of heat. In (Laguë and Swann 2016) afforestation in the 30°N –60°N latitudinal band, whatever the magnitude (increases in tree cover from 1.5 to 15.3 $Mkm^2$), induces increases in cross equatorial atmospheric energy transport and consequently a northward shift of the annual mean location of the ITCZ.

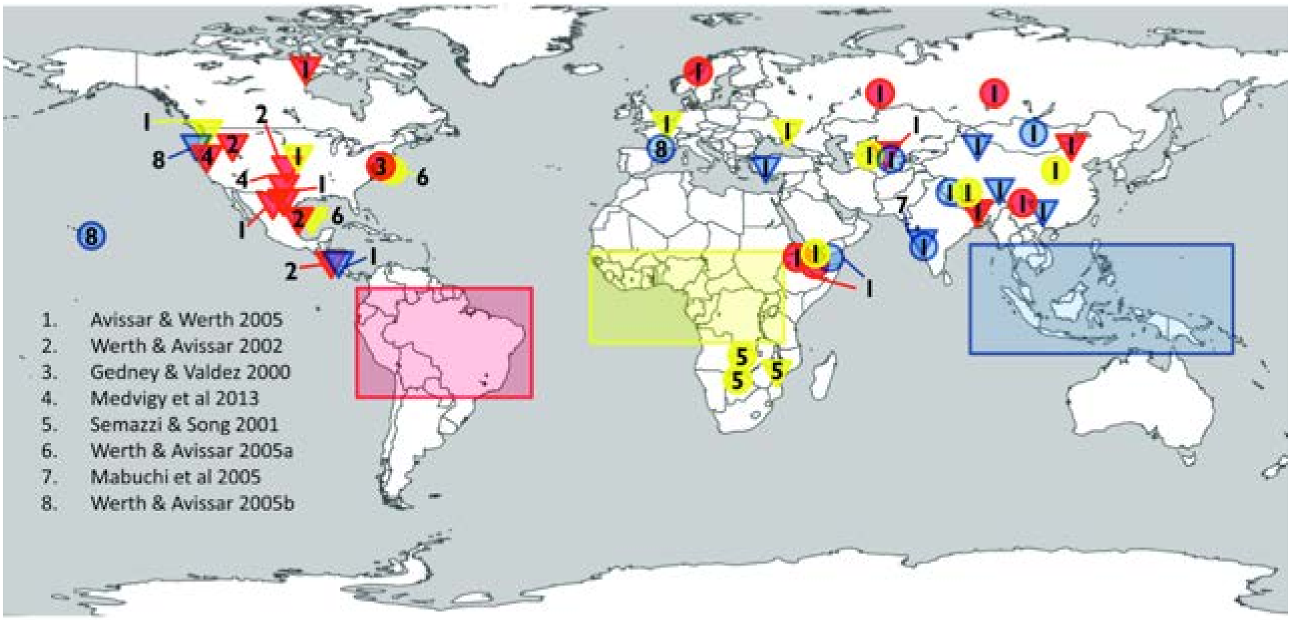


**Figure 2.30 Extra-tropical effects on precipitation due to deforestation in each of the three major tropical regions. Increasing (circles) and decreasing (triangles) precipitation result from complete deforestation of either Amazonia (red), Africa (yellow), or Southeast Asia (blue) as reviewed by (Lawrence and Vandecar 2015). Boxes indicate the area in which tropical forest was removed in each region. Numbers refer to the study from which the data were derived. Cited papers are the following (Avissar and Werth 2005; Gedney and Valdes 2000; Semazzi and Song 2001; Werth and Avissar 2002; Medvigy et al. 2013; Mabuchi et al. 2005; Werth and Avissar 2005; Werth 2005)**

De Vrese et al. (2016) demonstrated, using a global climate model, that irrigation in India affected regions as remote as eastern Africa through changes in the atmospheric transport of water vapour. At the onset of boreal spring (February to March) evapotranspiration is already large over irrigated crops and the resulting excess moisture in the atmosphere is transported south-westward by the low-level winds. This resulted in increases in precipitation as large as 1mm/day in the horn of Africa. Such finding, if robust, implies that if irrigation is to decrease in India, rainfall may decrease in eastern Africa where the consequences of drought are already disastrous.

The role of forests as providers of water vapor for the atmosphere and therefore as catalysts for increasing terrestrial precipitation has been questioned by (Ellison et al. 2017 and Layton and Ellison, 2016) based on a literature survey. They bring forward the potential of combined ‘small-scale’ afforestation and irrigation to boost the precipitation-recycling mechanism in the semi-arid region of Los Angeles California, and to provoke the growth of natural vegetation down-wind of the afforested area but up-wind of the ‘mountains’ (2.32).

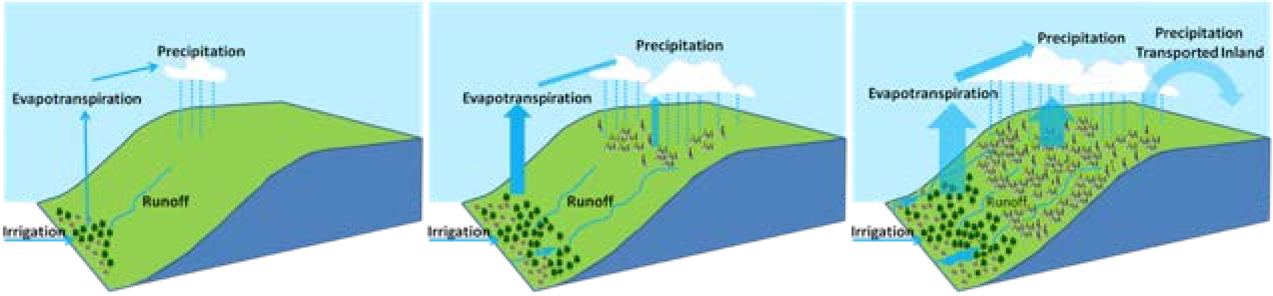


**Figure 2.31 Schematic illustration of how combined afforestation and irrigation in Los Angeles (California) area can influence downwind precipitation on mountainous areas, favor vegetation growth and feeds back to the afforested area via increased runoff (Layton and Ellison 2016). Areas of forests plantation and irrigation are located on the left panel, while downwind effects and feedbacks are illustrated in the middle and right panels**

Lorenz et al. (2016b) warned the community that existence of such teleconnections can be biased as a) the size of the imposed deforestation is often exaggerated in the modelling studies, b) the magnitude of the internal climatic variability is not always well accounted for in the statistical tests applied to estimate the significance of the changes simulated. Quesada et al. (2017a) however tackled those caveats using 1) realistic land use scenario (RCP8.5, projected reduction in deforestation rates compared to actual rates) and 2) several climate models to get intermodel robustness and significance. They concluded that large-scale teleconnections were at play.

In conclusion, land use changes do not just have local consequences (through biophysical, chemical or biological processes), but also affect adjacent or more remote areas. Any land use change has the potential to induce teleconnections within the climate system. Any action on land (to e.g., dampen global warming effects) thus needs to be anticipated not only for its local impacts but also for how it may affect other countries.

## 2.7 Climate consequences of land-based mitigation and adaptation response options

In this section we assess the climate consequences of land-based mitigation and adaptation response options. Climate mitigation in the land sector aims to reduce GHG emissions, enhance GHG removals and protect or enhance carbon stocks. Response options may also affect non-GHG climate forcers and biophysical properties related to climate. As some response options affect mitigation and adaptation simultaneously, we organise this section around integrated response options that affect climate regardless of whether they are applied for mitigation or for adaptation. Chapter 6 explores the interplay between mitigation and adaptation as well as other synergies and trade-offs. Response options specific to desertification, degradation and food security are described in more detail in Chapters 3, 4 and 5. A full list of response options across the SRCCL can be found in Chapter 6.

The Paris Agreement requires reaching a "balance between anthropogenic emissions by sources and removals by sinks of greenhouse gases". As some sector emission are hard to eliminate entirely (e.g., air transport, food production) this implies an ongoing need for options that result in a net removal of GHGs from the atmosphere and storage in living or dead plant material, soils, or in geological stores (Fuglestvedt et al. 2018). These are frequently referred to in the literature as carbon dioxide removal (CDR), Greenhouse Gas Removal (GGR) and negative emissions technologies (NETs). The response options discussed in his section that can contribute to CDR include afforestation/reforestation, forest management, soil carbon sequestration, biochar, Bioenergy with Carbon Capture and Storage (BECCS) and Enhanced Weathering.

Note that 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. Several assessments since AR5 have aggregated across the literature to assess the mitigation potential of individual response options building on a wealth of new literature. These are assessed in Section 2.7.1. The demand for, and potential of, land-`based mitigation under given climate mitigation targets is influenced by wider socioeconomic conditions, the interplay between different land-based mitigation options as well as with mitigation options in other sectors (such as energy or transport) and interaction with other sustainability goals. Thus, in Section 2.7.2 we look at modeled integrated assessment pathways highlighting the role of the land sector in contributing to specific mitigation pathways (2.7.2). At the end of this section we assess the potential impact of the Paris Agreement on the climate impacts of land response options. Chapter 7 further assesses policy and governance issues.

### 2.7.1 Assessing different land management response options and their potential consequences for climate mitigation

There are several key areas of uncertainty in assessing the potential consequences of land response options for climate mitigation. Estimates of what is considered to be "available" land for mitigation is often the starting point for assessments. It relies on interpretations of what land is "marginal", "degraded", etc. and may not take account of the existing uses and functions of that land such as subsidence farming, nomadic grazing or wildlife access. Most estimates prioritise land for food production, fiber and some degree of habitat protection before implementing mitigation options. Thus, estimates of mitigation potential are very sensitive to assumptions about future agricultural intensification (high crop yields, intensified pasture management and livestock production systems may decrease the need for agricultural expansion and in consequence free up more land for mitigation), use of fertilisers and irrigation, approaches to land protection or restoration, diet shifts away from high land intensity products, and reduction of waste.

The amount of land converted, the prior land cover, the amount of carbon in vegetation and soils, the rates of growth and decay, the location of activity and the required inputs (fertiliser, energy, etc.) alter the net GHG flux across the lifecycle of the response option. Varying definitions of land cover or activity (e.g., the definition of "forest"), time periods assessed, assumptions in population or demand for food and timber products, and carbon pools included (some only include aboveground biomass, while others include belowground biomass, dead wood, litter and soil, and even whether soil is peat) add further uncertainty. Some estimates include full life-cycle analysis from change in land cover, through inputs used, to transport and product disposal. Estimates are also sensitive to the counterfactual when considering "what would have happened otherwise" for example savings due to replacement of products-does bioenergy replace coal or gas or petrol, what would the land have been used for otherwise?

Changes in biophysical climate consequences are also sensitive to prior land cover. In general, there are significant differences in regional patterns of the responses to land use and management changes, and local biophysical climate effects following LUCs are more robust and larger in magnitude than global effects (Perugini et al. 2017).

Carbon stored in biomass and soils is at risk of future climate change, natural disturbances and future management changes for example harvesting of forests. Short-term losses due to harvesting or natural disturbance can be replaced by regrowth, as long as there is not permanent change in land cover for example planting crops after a forest fire. Thus, quantifying mitigation potential could be considered in relation to the change in the average long term carbon stock.

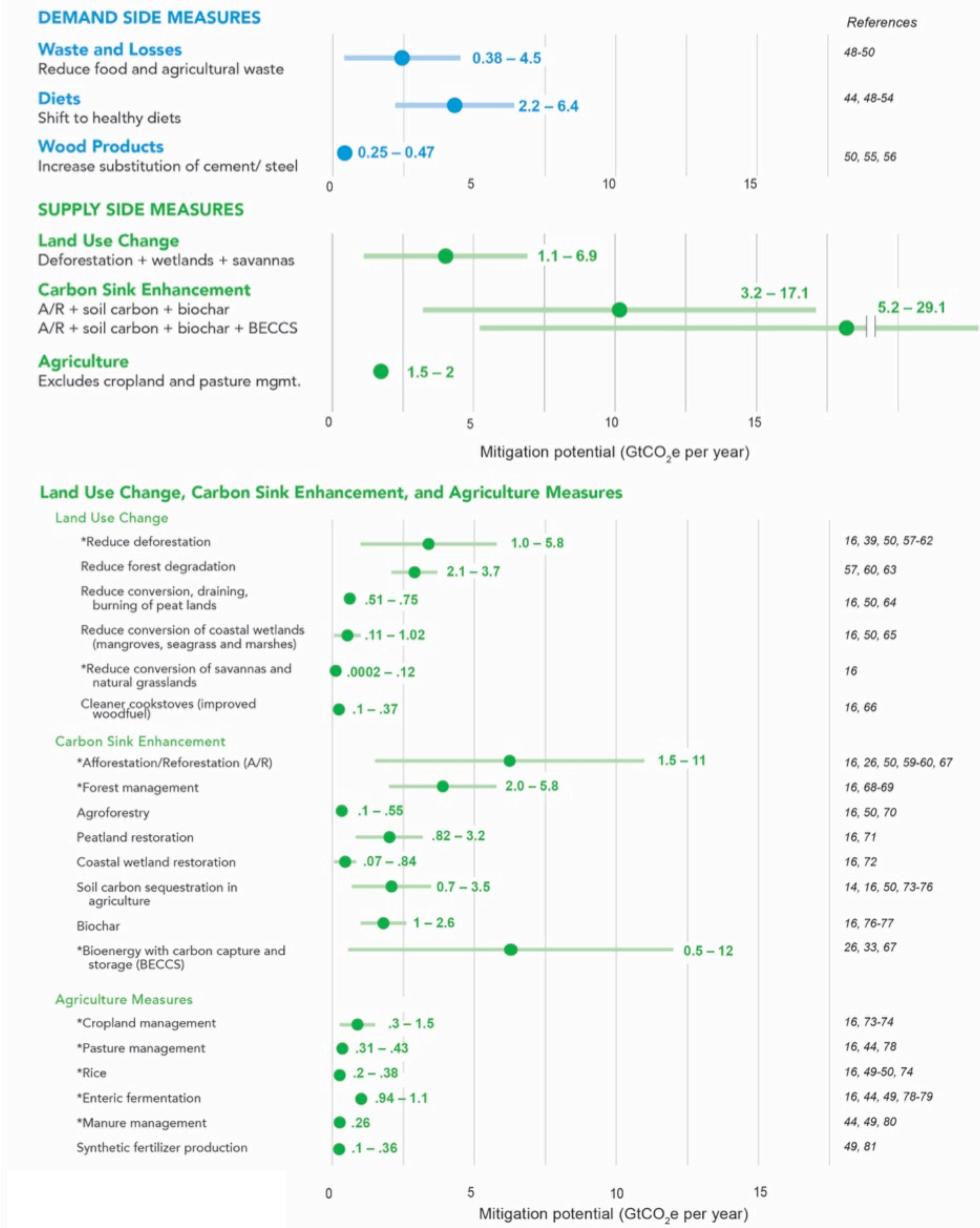


**Figure 2.32 Land-based mitigation potential by activity type, measured in $GtCO_2$-eq $yr^{-1}$ reflecting the full range of low to high estimates from literature (Roe et al.). Includes both technical (possible with current technologies) and economic (possible given economic constraints) mitigation potential. Only includes references that explicitly provide mitigation potential estimates in $CO_2$-eq $yr^{-1}$. Provides separate estimates for total supply-side and demand-side measures as these two categories are not additive. Elements of the analysis were designed to avoid potential double-counting of emissions reductions – the summed categories are highlighted in the supply-side and demand-side measures. Supply-side measures are activities that require a change in land use and/or management. Demand-side measures are activities that require a change in consumer behaviour**

When combining estimates from multiple bottom-up studies and sources, there are a range of methodologies reflected that may not be directly comparable or additive. Some of the studies assess the technical mitigation potential, or the amount of emissions reductions and carbon sequestration possible with current technologies without economic. Social and political constraints (Roe et al.). Some include biophysical or resource constraints (e.g., limits to yields, limits to natural forest conversion) to assess a “sustainable potential” (e.g., (Grosjean et al. 2018; Griscom et al. 2017), while some assess mitigation at different carbon prices – “economic mitigation potential” (e.g. (Smith et al. 2017; Griscom et al. 2017; Fuss et al. 2018). The multi-option assessment of Smith et al. (2017) focused on land-based CDR options and explored resource and sustainability issues. Griscom et al. (2017) focused on “Nature based solutions”, calculating potentials broken down by country after accounting for constraints around the production of food, fiber hand habitat for biological diversity.

They explored 2 carbon prices, 10 USD/Mg$CO_2$-eq to approximate existing prices, and 100 USD/Mg$CO_2$-eq as a maximum cost of emission reduction. In their three-part review, Minx et al. (2018), Fuss et al. (2018), and Nemet et al. (2018) provided a literature assessment of mitigation potentials, costs and side effects of seven negative emissions technologies, as well as analysed sustainable global potentials. Roe et al. (In press) conducted a bottom-up review of mitigation potential, assessing the range of technical and economic potential of 23 land-based activities in both supply and demand-side activities. They further combined their literature based assessment with integrated assessment model projections to highlight possible optimised future land based mitigation.

#### *2.7.1.1 Soil management response options*

Soils sequester C when the net inputs of organic matter (from plant litter, residues and manure) exceed losses of carbon from mineralisation, leaching and erosion. While the C stocks of natural land may decrease under agricultural use(Six et al. 2002; Paustian et al. 2016), optimisation of land management practices may allow these soils to become net C sinks. The rate of microbial SOM decomposition depends on the form of the inputs and existing pools; fresh OM (e.g., crop residues) has a typical decomposition rate several times greater than composted OM (Paustian et al. 2016). Formation of aggregates (Six et al. 2002) and complexes involving fine (clay and silt) soil particles (Feng et al. 2013; Hassink and Whitmore 1997) also affect SOC stabilisation and the capacity of soils to stabilise additional organic matter inputs (Beare et al. 2014), meaning not all will sequester C at the same rate. Formation of calcium carbonates in the soil provides a permanent sink for mineralised organic C (Manning 2008; Beerling et al. 2018) (see also 2.7.1.7). Potential for soil carbon sequestration therefore varies considerably, depending on prior and current land management approaches, soil type, resource availability, environmental conditions microbial and fungi composition and nutrient availability among other (Smith and Dukes 2013; Palm et al. 2014; Lal 2016). Soils are a finite sink, leading to uncertainty around their ability to provide ongoing long-term GHG removals. C sequestration rates may potentially decline to negligible levels over as little as a couple of decades as soils reach C saturation (West et al. 2004; Smith and Dukes 2013).

Land management practices to increase C include improved rotations with deeper rooting cultivars, reduced- or zero-tillage cultivation, addition of organic materials, and agroforestry(Lal 2011; Smith et al. 2008; Lorenz and Pitman 2014; Lal 2016; Vermeulen et al. 2012a; de Rouw et al. 2010). There is *high agreement and medium evidence* that adoption of green manure cover crops, while increasing cropping frequency or diversity, helps sequestering SOC (Poeplau and Don 2015; Mazzoncini et al. 2011; Luo et al. 2010). Implementation of agroforestry systems on arable land also has potential sequester SOC (Nair et al. 2009; Corbeels et al. 2018; Lorenz and Lal 2014) (Section 2.7.1.4, Chapter 5). There is *medium evidence* that conservation agriculture, i.e. the simultaneous adoption of minimum tillage, (cover) crop residue retention and associated soil surface coverage, and crop rotations, can contribute to long-term SOC sequestration; both, positive (Powlson et al. 2016; Zhang et al. 2014) and inconclusive cases (Cheesman et al. 2016; Palm et al. 2014; Govaerts et al. 2009), have been published. The efficacy of reduced and zero-till practices is highly context-specific; many studies (e.g. Paustian et al., 2000; Six et al., 2004; van Kessel et al., 2013) demonstrate increased carbon storage, while others show the opposite effect (Sisti et al. 2004; Álvaro-Fuentes et al. 2008; Christopher et al. 2009). Meta-analyses (Haddaway et al. 2017; Luo et al. 2010) also show mixed responses, and the lack of robust comparisons of soils on an equivalent mass basis continues to be an problem for credible estimates (Wendt and Hauser 2013; Powlson et al. 2011; Powlson et al. 2014).

The technical potential for soil carbon sequestration globally has been estimated between 1.1 and 11.4 $GtCO_2$ pa, with more conservative estimates between 3.6 and 6.9 $GtCO_2$ $yr^{-1}$ (Conant et al. 2011; Minasny and McBratney 2018; Lal 2011, 2016) *(Robust evidence, medium agreement)*. The soil mitigation potential in agricultural systems through conservation agriculture practices (including reduced and zero tillage, crop residue management, use of perennials or deeper rooted cultivars, organic amendment and fire management), and pasture management (including managing stocking rates, timing and rotation of livestock, higher productivity grass species or legumes, and nutrient management) to be 0.7–3.5 Gt $CO_2$ $yr^{-1}$ (estimated from (Roe et al.) based on (Frank et al. 2017a; Griscom et al. 2017; Smith 2016; Paustian et al. 2016; Sanderman et al. 2017; Bernoux and Paustian 2014). Management of soil erosion could switch from net emissions of 1.36– 3.67 to net removals of 0.44–3.67 $GtCO_2$ $yr^{-1}$ *(robust evidence, kow agreement)* (Jacinthe and Lal 2001; Lal et al. 2004; Stallard 1998; Smith et al. 2001; Van Oost et al. 2007). Assuming unit costs limited to between 5 and 25 USD per $tCO_2$ yields estimates of global carbon emission mitigation potentials of soil carbon sequestration between 1.5 and 2.6 $GtCO_2$ $yr^{-1}$ for a period of 10–20 years (Smith et al. 2016c, 2008).

Soil carbon management does not have specific land requirements as it happens in situ on existing land use, but it requires changes in practices that may affect other aspects of land management. Soil carbon management interacts with $N_2O$ emissions(Poeplau and Don 2015; Paustian et al. 2016). For example, (Li et al. 2005) estimate that the management strategies required to increase C sequestration (reduced tillage, crop residue, and manure recycling) would increase $N_2O$ emissions significantly, offsetting 75–310% of the C sequestered.

### *2.7.1.2 Cropland, grassland and livestock management response options*

Reducing emissions intensity from enteric fermentation, manure management, rice fields and fertiliser production has a total mitigation potential of 1.5–2.1 Gt $CO_2$-eq $yr^{-1}$ *(robust evidence, medium agreement)* (Zhang et al. 2013b; Hristov et al. 2013; Dickie et al. 2014; Herrero et al. 2016; Henderson et al. 2015; Paustian et al. 2016; Griscom et al. 2017; Hawken 2017a). The mitigation potential of cropland management is 0.3–1.5 Gt $CO_2$-eq $yr^{-1}$ (Griscom et al. 2017; Paustian et al. 2016) and pasture management is 0.3–0.4. Gt $CO_2$-eq $yr^{-1}$ *(robust evidence, medium agreement)*(Anderson and Peters 2016; Herrero et al. 2016; Henderson et al. 2015; Paustian et al. 2016; Griscom et al. 2017).

Since agriculture accounts for 56% of methane emissions, and 27% of all potent short-lived gases (Sections 2.4 and 2.5), measures addressing enteric fermentation, manure management and rice $CH_4$ emissions have strong potential to reduce global warming effects in the short-term throughout many regions (Montzka et al. 2011; Collins et al. 2018; Hayman et al.). Enteric fermentation is responsible for over 40% of direct agricultural emissions with beef and dairy cattle accounting for approximately 65% (Herrero et al. 2016). Manure from livestock cause both nitrous oxide and methane emissions, and account for roughly one quarter of direct agricultural GHG emissions (Dickie et al. 2014). The three main measures to reduce enteric fermentation include improved diets (higher quality, more digestible livestock feed), supplements and additives (reduce methane by changing the microbiology of the rumen), and animal management and breeding (improve husbandry practices and genetics (Zhang et al. 2013b; Hristov et al. 2013; Dickie et al. 2014; Herrero et al. 2016; Henderson et al. 2015; Paustian et al. 2016; Griscom et al. 2017; Hawken 2017a). Measures to manage manure include anaerobic digestion for energy use, composting as a nutrient source, reducing storage time, and changing livestock diets.

Reducing emissions from rice production through improved water management (periodic draining of flooded fields to reduce methane emissions from anaerobic decomposition), and straw residue management (apply in dry conditions instead of on flooded fields, avoid burning to reduce methane and nitrous oxide emissions) has the potential to mitigate up to 60% of emissions (Hussain et al. 2015), or 0.2–0.38 Gt $CO_2$ $yr^{-1}$ *(robust evidence, medium agreement)* (Dickie et al. 2014; Paustian et al. 2016; Griscom et al. 2017; Davin et al. 2014; WANG et al. 2018). Synthetic fertiliser production is a major source of GHG emissions and air pollution as it requires a large amount of energy to produce, and uses fossil fuels (natural gas or coal) as feedstocks. Improvements in industrial efficiency are typically cost effective, would improve the productivity of the sector, reduce pollution, and have the potential to mitigate emissions (Zhang et al. 2013b; Dickie et al. 2014).The emissions intensity of agriculture can be improved by more efficient use of inputs (nutrient and organic amendments, water, energy) or increasing inputs (used in a an efficient way), including use of improved varieties/breeds, and mechanisation. A variety of farming practices can be adopted that optimise density, rotations and precision of inputs. Intensification could lower agricultural expansion rates and hence GHG emissions from land use change but may produce an increase in fertiliser use and other agrochemicals which may increase emissions and pollution (Garnett et al. 2013). Further, more efficient production methods can reduce costs and increase yields, and therefore, may encourage farmers to further increase production and expand land use (deforest) (Lambin and Meyfroidt 2011). A broad discussion on these supply side mitigation options and their cobenifits are provided in Chapter 5, Section 5.5.1.

*2.7.1.2.1 Demand-side management in the food sector (diet change, waste reduction)*

Demand-side management has the potential for climate change mitigation via reducing emissions from production, switching to consumption of less emission intensive commodities, and making land available for carbon dioxide removal (see Chapter 5 Section 5.5.2). Reducing food losses and waste increases the overall efficiency of food value chains (less land and inputs needed) along the entire supply chain (Porter et al. 2016; Hiç et al. 2016). Demand-side measures have the potential to significantly mitigate emissions of 2.83–11.37 Gt $CO_2$-eq $yr^{-1}$ from reductions in food loss and waste (food wastage), changes in diets, and the substitution of wood for cement and steel in construction (see below). Approximately 55% of the upper bound of this estimate comes from changes in diet and the other 40% comes from reductions in food wastage (Roe et al.).

Shifting to healthier diets and away from emissions-intensive foods like beef delivers a significant mitigation potential of 2.2–6.4 Gt $CO_2$-eq $yr^{-1}$ (Bajželj et al. 2014; Dickie et al. 2014; Herrero et al. 2016; Hawken 2017; Springmann et al. 2016; Tilman and Clark 2014; Hedenus et al. 2014; Stehfest et al. 2009). Countries with the highest overall and projected beef consumption include predominantly developed and emerging countries: USA, EU, China, Brazil, Argentina, Russia. A recent study concluded that reducing beef consumption (decrease in the USA by 50%, Brazil by 25% and stabilisation in China) could provide a 12% mitigation of livestock emissions by 2030 solely from avoided enteric fermentation and manure emissions (Haupt 2017). In addition to mitigation gains, decreasing meat consumption, primarily of ruminants, and reducing wastes further reduces water use, soil degradation, pressure on forests, land used for feed, and manure and pollution into water systems and be important measures for adaptation including increasing food supply (Tilman and Clark 2014) (see chapters 5 and 6).

*2.7.1.2.2 Forest-related response options*

Mitigation and adaptation response options in forests and other wooded lands that have climate consequences are already widely practiced throughout the world. They are thus often used in modeled scenarios as immediately available mitigation methods, while other technologies are being developed towards large-scale deployment.

The potential for reducing and/or halting deforestation and degradation range between 1.2–5.8 Gt $CO_2$-eq $yr^{-1}$, with the higher figure representing a complete halting of land use conversion in forests and peatlands and accounting for biomass and soil *(robust evidence, high agreement*(Griscom et al. 2017; Busch and Engelmann 2017; Baccini et al. 2017; Carter et al. 2015; Houghton 2013; Houghton and Nassikas 2018b; Smith and Dukes 2013; Zarin et al. 2016). Reduced deforestation and degradation includes conservation of existing carbon pools in vegetation and soil through protection in reserves, controlling disturbances such as fire and pest outbreaks, and changing management practices. The mitigation estimates represent biophysical and technical potential (higher ranges) and economic and feasible mitigation potential (lower ranges). Differences in estimates also stem from varying land cover definitions, time periods assessed, and carbon pools included (most lower estimates only include aboveground biomass, and most higher estimates include all five IPCC carbon pools: aboveground, belowground, dead wood, litter, soil, and peat). When deforestation and degradation are halted, some gross emissions would continue due to ongoing decomposition of residues and soil carbon, and it may take many decades to fully recover the biomass initially present in native ecosystems. (Houghton and Nassikas 2018b) have estimated a cumulative sequestration potential of 439 $GtCO_2$ $yr^{-1}$ between 2016 and 2100 if deforestation and wood harvest were stopped and secondary forests were allowed to recover.

Afforestation/Reforestation (A/R) can increase carbon sequestration in both vegetation and soils by 1.5–11 Gt $CO_2$ $yr^{-1}$ *(robust evidence, medium agreement)*(Houghton 2013; Lenton 2014; Smith et al. 2016a; Houghton and Nassikas 2017; Griscom et al. 2017). The lower estimate takes into consideration the competition for land use in agriculture, and the high estimate allows all secondary forests and shifting cultivation fallow areas to naturally regenerate. The most recent mitigation potential estimates for A/R provide "realistic" figures of 4.04 $GtCO_2$ $yr^{-1}$ by 2100 (Smith et al. 2016b) by averaging model results that factor in deployment costs in a 2°C scenario, and 3.04 $GtCO_2$ $yr^{-1}$ by 2030 (Griscom et al. 2017) by considering spatially explicit environmental and social constraints as well as economic constraints of <\$100 per $tCO_2$. Future sequestration potentials for afforestation, reforestation and forest management modelled in line with the 1.5°C and 2°C scenarios, were found to increase to around 4 $GtCO_2$-eq $yr^{-1}$ by 2070 (Roe et al.) (See also 2.7.2). A/R takes some time for full carbon removal to be achieved as the forest grows, with net uptake of carbon slowing as forests reach maturity. Significant amounts of land are required to achieve substantial $CO_2$ capture, with land intensity of afforestation and reforestation estimated at 0.29 ha $tCeq^{-1}$ $yr^{-1}$ (Smith et al. 2016a). Boysen et al. (2017) estimated that to sequester about 100 GtC by 2100 would require 1300 Mha of abandoned cropland and pastures.

Forest management has the potential to mitigate 2-5.8 Gt $CO_2$ $yr^{-1}$ in 2030, although the upper estimate also includes some A/R *(robust evidence, medium agreement)*(Nabuurs et al. 2017; Bosello et al. 2009; Sasaki et al. 2016; Griscom et al. 2017). Forest management can alter productivity, turnover rates, harvest rates and carbon in wood products, with the balance resulting in long-term increases or decreases in carbon stocks (Nabuurs et al. 2007; Lemprière et al. 2013; Kurz et al. 2016) (see also Chapter 4, Section 4.7.6). Specific mitigation practices include extending rotation cycles between harvests, reducing damage to remaining trees when harvesting, reducing logging waste, implementing soil conservation practices, and using wood more efficiently. Fertilisation may enhance productivity but would increase $N_2O$ emissions. Carbon removal from the atmosphere occurs at faster rates in young to medium aged forests and declines thereafter such that older forest stands have smaller carbon removals but larger stocks (Tang et al. 2014). The trade-off between maximising forest C stocks and maximising ecosystem C sinks leads to uncertainty about optimum management strategies to achieve negative emissions (Keith et al. 2014; Kurz et al. 2016). Timber can be harvested and used for bioenergy substituting for fossil fuels (with or without carbon capture and storage) (Section 2.7.1.5), long-lived products such as timber (See below), or buried as biochar (Section 2.7.1.1), enabling areas of land to be used continuously for mitigation providing harvest is followed by regrowth.

Global mitigation potential from increasing the demand of timber products to replace construction materials range from 0.25–0.48 $GtCO_2$ (Roe et al.) with the low estimate assuming a material substitution effect of 0.28 tC $m^{-3}$ of final wood product, and a roundwood volume of 0.9 billion $m^3$ annually (based on 2000 level demand) (Pekka Kauppi 2001) and the high estimate assuming 40% of global wood products were used for construction (Miner 2010) (*low evidence, medium agreement)*. Using harvested carbon in long-lived products (e.g., for construction) can represent a store that can sometimes be from decades to over a century while the wood can also substitute for intensive building materials, avoiding emissions from the production of concrete and steel(Sathre and O'Connor 2010; Smyth et al. 2017; Nabuurs et al. 2007; Lemprière et al. 2013). The harvest of carbon and storage in products affects the net carbon balance of the forest sector, with the aim of sustainable forest management strategies being to optimise carbon stocks and use of harvested products to generate sustained mitigation benefits (Nabuurs et al. 2007a; Cherubini et al. 2012; Earles et al. 2012; Marland et al. 2010). The displacement factor, or the substitution benefit in $CO_2$, when wood is used instead of another material estimated in the literature ranges range from -2.3 to 15 $tCO_2$ of emission reduction per tC in wood product (Sathre and O'Connor 2010).

The mitigation potential from agroforestry ranges between 0.55–1.04 Gt $CO_2$ $yr^1$, (*medium evidence, high agreemen)*(Gharajehdaghipour et al. 2016; Griscom et al. 2017; Zomer et al. 2016). Agroforestry is a land management system that combines woody biomass (e.g., trees or shrubs) with crops and/or livestock). Zomer et al. (2016) reported that the trees in agroforestry landscapes had increased carbon stock by 7.33 $GtCO_2$ between 2000 and 2010, or 0.7 $GtCO_2$ $yr^{-1}$. It can increase carbon stocks in vegetation and soils and reduce emission from CH4 and N2O(Mutuo et al. 2005; Rosenstock et al. 2014; Montagnini and Nair 2004). Mbow et al. (2014) suggest that agro-forestry systems have the potential to contribute to 0.5 to 3.1 Mg C $ha^{-1}$ $yr^{-1}$ of carbon sequestration in Africa. For more details see Chapter 5, Section 5.1.1.3.

Biophysical effects of forest response options are variable depending on the location and scale of activity (Nobre et al. 2009; Davin and de Noblet-Ducoudre 2010) (Section 2.6). In tropical areas, ground and satellite-based observational studies and modelling studies consistently indicate that deforestation causes a net biophysical warming where trees are removed and around, while globally a cooling is generally simulated when oceans are interacting (Bala et al. 2007; Bathiany et al. 2010; Devaraju et al. 2018; Kendra Gotangco Castillo and Gurney 2013; Lawrence and Vandecar 2015; Perugini et al. 2017; Longobardi et al. 2016; Wang et al. 2014b) (Section 2.6). Therefore, avoided deforestation or afforestation in the tropics contributes to climate mitigation through both biogeochemical and biophysical effects. It also maintains rainfall recycling to some extent though not always in areas afforested as discussed in Section 2.6 (increases in rainfall may in neighbouring regions). In contrast, in higher latitude boreal areas observational and modelling studies show that afforestation and reforestation lead to local and global warming effects, particularly in snow covered regions in the winter as the albedo is lower for forests than bare snow. , (Bathiany et al. 2010; Dass et al. 2013; Devaraju et al. 2018; Ganopolski et al. 2001; Snyder et al. 2004; West et al. 2011; Arora and Montenegro 2011). However global effects are quite small when compared to fossil $CO_2$-induced changes in temperature. Thus the biophysical effects run counter to the GHG effects in terms of climate mitigation at the local to regional scale with implications for adaptation (Section 2.6).

The biophysical impacts of forest area change in mid-latitudes are complex, because of the opposite effects on climate from the typical changes in albedo and surface evapotranspiration (Ma et al. 2017b; Devaraju et al. 2015; West et al. 2011; Snyder et al. 2004; Dümenil Gates and Ließ 2001). Recent studies however report that there is a strong seasonal difference at those latitudes with deforestation-induced warming during the growing season (late spring, summer and early fall) and deforestation-induced cooling outside the growing season (winter and early spring) (Ahlswede and Thomas 2017; Anderson-Teixeira et al. 2012; Anderson et al. 2011; Bright et al. 2017; Chen et al. 2012; Gálos et al. 2011, 2013; Strandberg et al. 2018; Wickham et al. 2013; Zhao and Jackson 2014). Below 35°N, afforestation and reforestation are considered to decrease near-surface temperature because the warming effect of decreased albedo is weaker than cooling effect of increased latent heat, roughness length and rooting depth (Pielke et al. 2007; Peng et al. 2014). Therefore for recent trends of afforestation in mid-latitude regions such as China, Russia, USA and Europe (Liu et al. 2015b) the biophysical effects support the climate mitigation mediated through biogeochemical effects during the vegetation growing period, although less strongly than in tropical areas (Perugini et al. 2017). It does not support climate mitigation when there is snow on the ground,

Forest cover also affects climate through reactive gasses and aersools (see Section 2.5) (Unger 2014b) *[placeholder - to be expanded]*.

*2.7.1.2.3 Wetland, peatland and coastal habitat management*

Protection and restoration of wetlands, peatlands and coastal habitats (such as mangrove forests, salt marshes and seagrass meadows) reduces net carbon loss (primarily from sediment/soils) and provides continue or enhanced natural $CO_2$ removal. Reducing annual emissions from peatland conversion, draining and burning would mitigate 0.51–0.75 Gt $CO_2$-eq $yr^{-1}$ (Hooijer et al. 2010; Griscom et al. 2017). Approximately 1 Gt $CO_2$ $yr^{-1}$ can be mitigated if 30% of the 65 Mha of drained peatlands were rewetted to stop continued emissions from carbon oxidation, and about 3.2 Gt $CO_2$ $yr^{-1}$ if all ongoing $CO_2$ emissions from continued peat oxidation were ceased (Joosten and Couwenberg 2008). Griscom et al. (2017) estimate 0.81 Gt $CO_2$ $yr^{-1}$ as a feasible target (<USD100 $tCO_2e^{-1}$) for rewetting and biomass enhancement. Wetland drainage and rewetting was included as a flux category under the second commitment Period of the Kyoto protocol, with significant management knowledge gained over the last decade (IPCC 2013). However there are high uncertainties as to the carbon storage and flux rates, in particular the balance between $CH_4$ sources and $CO_2$ sinks (Spencer et al. 2016). Peatlands are highly vulnerable to changes in temperature and precipitation under climate change (Clark et al. 2010; Gallego-Sala and et 2010).

The climate mitigation potential of mangrove forests is considered in Chapter 5 of the IPCC Special Report on the Ocean, Cryosphere and Climate Change, in a wider 'blue carbon' context. There is *high confidence* that climatically-significant carbon losses from mangrove land use change (Section 2.4) can be prevented, primarily by increased enforcement of existing regulatory measures (Miteva et al. 2015; Howard et al. 2017; Herr et al. 2017). Commitments to strengthen mangrove protection and conservation have been made in many Nationally Determined Contributions (NDCs) to the Paris Agreement (Gallo et al. 2017). The ongoing benefits provided by mangroves as a natural carbon sink can also be nationally-important for Small Island Developing States (SIDS), based on estimates of high carbon sequestration rates per unit area (McLeod et al. 2011; Duarte et al. 2013; Duarte 2017) although global totals are small compared to other ocean and land-based mitigation options (Gattuso et al. 2018; Griscom et al. 2017). Reducing the conversion of coastal wetlands (mangroves, seagrass and marshes) would realise mitigation of 0.11–1.02 Gt $CO_2$-eq $yr^{-1}$ of emissions (Pendleton et al. 2012; Griscom et al. 2017). Mangrove restoration can mitigate the release of 0.07 Gt $CO_2$ $yr^{-1}$ through rewetting (Crooks et al. 2011) and 0.84 Gt $CO_2$ $yr^{-1}$ from biomass and soil enhancement (Griscom et al. 2017). There is only *medium confidence* in the effectiveness of enhanced carbon uptake using mangroves, due to the many uncertainties regarding the response of mangroves to future climate change (Jennerjahn et al. 2017); dynamic changes in distributions (Kelleway et al. 2017) and other local-scale factors affecting long-term sequestration and climatic benefits (e.g., methane release; Dutta et al. 2017).

##### *2.7.1.2.4 Biochar*

Biochar is produced by thermal decomposition of biomass in the absence of oxygen (pyrolysis) into a stable, long-lived product like charcoal that is relatively resistant to decomposition (Lehmann et al. 2015) and which can stabilise organic matter added to soil (Weng et al. 2017) (Although charcoal has been used traditionally by many cultures as a soil amendment, "modern biochar", produced in facilities that control emissions, is not widely used. A global analysis of technical potential, in which biomass supply constraints were applied to protect against food insecurity, loss of habitat and land degradation, estimated *technical potential* abatement of 3.7–6.6 $GtCO_2$-eq $yr^{-1}$ (including 2.6–4.6 $GtCO_2$-eq $yr^{-1}$ carbon stabilisation), with theoretical potential to reduce total emissions over the course of the century by 240–475 $GtCO_2$-eq ((Woolf et al. 2010). (Fuss et al. 2018) propose a range of 0.5–2 $GtCO_2$-eq as the *sustainable potential* for negative emissions through biochar. (Griscom et al. 2017) suggest a potential of 1.0 $GtCO_2$ $yr^{-1}$ based on available residues. Biochar is discussed in more detail in Chapter 4.

##### *2.7.1.2.5 Biomass provision for bioenergy and BECCS*

Mitigation potential from BECCS is estimated to be between 0.5 and 12.0 GtCO2 $yr^{-1}$ *(robust evidence, medium agreement)* (Smith et al. 2016c; Fuss et al. 2018; Lenton 2014). The overall mitigation potential of bioenergy and BECCS depends on case-specific considerations related to life-cycle GHG fluxes and land/climate interactions, such as the feedstock used (e.g., residues, dedicated crops, algae), prior land use (e.g., degraded lands or high carbon forest lands), GHG fluxes due to biomass production (productivity, soil carbon dynamics, and fertiliser use), the bioenergy pathway and product (e.g., wood pellets, ethanol), the type of fuel being substituted (e.g., coal or oil) *(high evidence, high agreement).*

*Life-cycle emissions:* Direct life-cycle emissions of most modern bioenergy alternatives constitute net savings in comparison to fossil fuels, providing they do not result on conversion of ecosystems high in carbon (discussed below) *(high evidence, medium agreement)* (Chum et al., 2011; Creutzig et al., 2015). The steps required to cultivate, harvest, transport, process and use biomass fuels generate emissions of GHGs and other climate pollutants (Staples et al., 2017). The magnitude of these impacts largely depends on the type of biomass, transportation distances, conversion technologies, applications, and location (Chum et al., 2011; Creutzig et al., 2015; Muñoz et al., 2014; Müller-Langer et al., 2014). They are usually lower for advanced bioenergy and biofuels from forest-based resources and agricultural residues, whereas for dedicated bioenergy crops the agricultural phase can be more energy and GHG intensive (Chum et al., 2011; Gerbrandt et al., 2016). For example, short rotation coppice or perennial grasses such as switchgrass and miscanthus often requires application of N-containing fertilisers to achieve high yields, thereby enhancing soil emissions of $N_2O$, although at usually lower rate than conventional food crops (Lai et al., 2017; Oates et al., 2016; Robertson et al., 2017; Rowe et al., 2016). At a global level, an optimal reduction of life-cycle GHG emissions from biofuels relative to fossil fuels under a 2°C mitigation pathway ranges from 18% to 61% in 2050 (between 5 and 16 $GtCO_2$ $yr^{-1}$), depending on different assumptions about biomass and land availability (Staples et al., 2017).

*Land use, direct land use change and carbon payback times:* The net greenhouse gas balance of bioenergy systems can be positively or negatively affected by land use and direct land use changes, depending on previous land use and management practice *(high evidence, high agreement)* (Berndes et al., 2013; Elshout et al., 2015). In general, large-scale deployment of bioenergy requires significant amounts of land to provide biomass feedstocks and thus increases land competition with risks for conversion of carbon-rich ecosystems, whereas low levels of bioenergy deployment and the use of residues reduce these effects (Creutzig et al., 2015; Vaughan et al., 2018). Estimates of marginal lands currently considered "available" for bioenergy production range from 385 to 1100 Mha, depending on land class definitions, soil conditions, and land mapping method, although noting that even these lands are likely under some degree of use (Cai et al., 2011; Campbell et al., 2008; Lewis and Kelly, 2014). These differences lead to a wide range of estimates with low agreement on the quantitative mitigation potential.

Establishing bioenergy crops in previous high carbon forestland or peatland in the tropics or high latitudes results in high emissions of carbon that may take up to a century to be re-paid in terms of net $CO_2$ emission savings from replacing fossil fuels (Elshout et al., 2015). For example, carbon payback times associated with a 1.5 degree IAM scenario varied from insignificant when replacing agricultural crops in temperate areas, through 10 to 100+ years replacing tropical forests, and over 100 years due to loss of soil carbon in high latitude peatland forests (See Figure 2.33) (Harper et al., 2018), although the result is very sensitive to bioenergy crop yield estimates. On the other hand, short rotation coppice species or perennial grasses established on former cropland or degraded land typically accumulate carbon in soils thanks to their deep root system at rates up to 0.7 tC $ha^{-1}$ $yr^{-1}$ (Don et al., 2012). In the case of bioenergy from managed forests, the carbon payback time is controversial (Berndes et al., 2013; Cherubini et al., 2016; Cintas et al., 2017; Hudiburg et al., 2011). In some studies, it ranges from a few years to more than a century depending on plant species, local climate, residue management, or fossil fuel displaced (Guest et al., 2013; Lamers and Junginger, 2013; Ortiz et al., 2016; Ter-Mikaelian et al., 2015). Other studies arrive at opposing conclusions when accounting from a forest owner's perspective and their expectations about market development for bioenergy and other wood-based products (Berndes et al., 2013; Cintas et al., 2017). In terms of climate system response, whereas $CO_2$ emissions from fossil fuels cause a nearly irreversible warming (Eby et al., 2009; Solomon et al., 2009), the forcing from bioenergy systems is temporary and less relevant to long-term temperature stabilisation, provided the biomass is regrown (Cherubini et al., 2014; Jones et al., 2013; Mackey et al., 2013).

*Indirect land use change (iLUC):* Attribution of emissions from iLUC to bioenergy (mostly crop-based biofuels) risk undermining its net climate change mitigation benefits *(low evidence, low agreement).* Bioenergy from dedicated crops are in some cases held responsible for GHG emissions resulting from indirect land use changes, that is the bioenergy activity may lead to displacement of agricultural or forest activities into other locations, driven by market-mediated effects. While this is the case for any mitigation options requiring land (e.g., afforestation or avoided deforestation), it is a concern most commonly raised in relation to bioenergy. The term "indirect" is used to capture leakage effects occurring in other sectors or regions, and indirect emissions are only specified in the context of a specific product, sector or region, because at global levels all land use emissions are direct. These indirect emissions are potentially more significant for crop-based feedstocks such as corn, wheat and soybean, than for advanced biofuels from lignocellulosic materials (Ahlgren and Di Lucia, 2014; Chum et al., 2011; Valin et al., 2015; Wicke et al., 2012). Estimates of emissions from indirect land use change are inherently uncertain and highly dependent on modelling assumptions, such as supply/demand elasticities, productivity estimates, incorporation or exclusion of emission credits for coproducts, and are widely debated in the scientific community (Finkbeiner, 2014; Kim et al., 2014; Rajagopal and Plevin, 2013; Zilberman, 2017) Wise et al. 2015). For example, iLUC values for corn bioethanol were originally estimated at 104 g $CO_2$ $MJ^{-1}$ fuel (Searchinger et al., 2008), but more recent estimates converge towards about 20 g $CO_2$ $MJ^{-1}$ fuel, or even lower (Ahlgren and Di Lucia, 2014). *[Placeholder for a figure illustrating effects of BECCS on C fluxes and biophysical feedbacks]*.

*Biophysical effects:* Bioenergy deployment can have large effects on regional climate, with the direction and magnitude of the impact depending on the type of bioenergy crop, previous land use and seasonality (*low evidence, high agreement*). Bioenergy-induced land cover changes exert biophysical mechanisms that can substantially influence the climate from local to regional scales. For example, land use changes due to future biofuel scenarios over the first half of the 21st century show a nearly neutral effect on surface temperature, as warming from GHG emissions and cooling from biophysical effects are nearly offsetting each other when averaged at global levels, although there are significant seasonal and regional differences (Hallgren et al., 2013). In particular, the switch from annual crops to perennial bioenergy plantations like *Miscanthus* in the US is an option to achieve synergies between climate change mitigation and local climate benefits, as it leads to regional cooling due to local increases in evapotranspiration and albedo (Georgescu et al., 2011; Harding et al., 2016). Perennial bioenergy crop expansion over suitable abandoned and degraded farmlands to avoid competition with existing food cropping systems causes near-surface cooling up to 5°C during the growing season in large portions of the central USA (Wang et al., 2017). Similarly, growing sugarcane at expenses of existing cropland in Brazil cools down the local surface during daytime conditions by about -0.8°C to -1°C, but warmer conditions occurs if sugar cane is deployed at the expense of natural vegetation (Brazilian Cerrado) (Loarie et al. 2011). Seasonal variations are significant, and they can be masked if annual means only are considered. In general, bioenergy crops induce a cooling of ambient air during the growing season, but after harvest the decrease in evapotranspiration can induce warming (Georgescu et al., 2013; Harding et al., 2016; Wang et al., 2017). Neglecting biophysical effects from establishment of bioenergy crops can underestimate their mitigation potential (Zhu et al., 2017), and prevent the identification and management of potential synergies with climate change adaptation at local and regional scales.

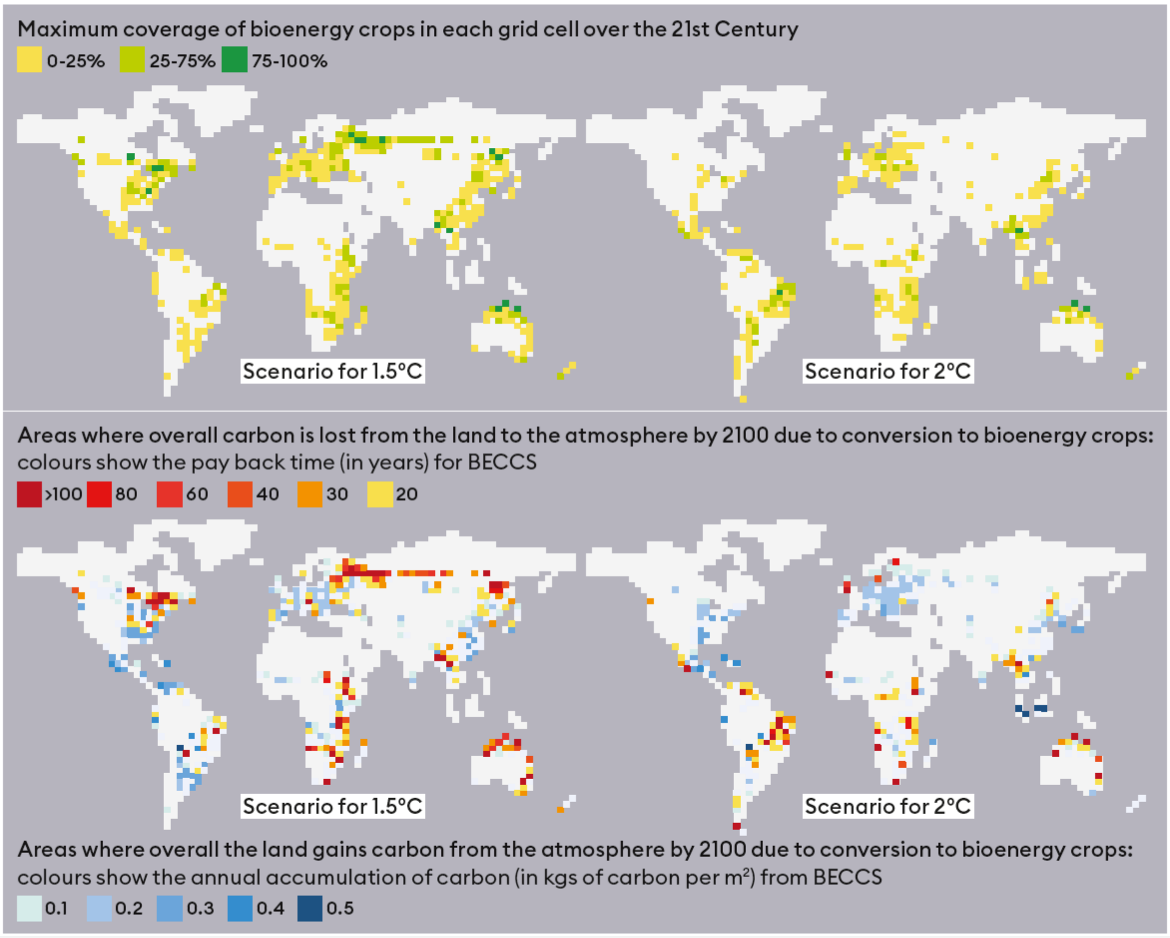


**Figure 2.33 Location of bioenergy crops in 1.5 and 2 degree scenarios and associated carbon payback times. Adapted from (Harper et al. 2018). The scenarios were produced by the IMAGE Integrated Assessment Model (Stehfest et al. 2014) using a central mitigation pathway (Shared Socioeconomic Pathway 2, SSP2-RCP1.9 or IM1.9 and SSP2-RCP2.6, or IM2.6). Land-based mitigation options are part of the overall mitigation portfolio, while maintaining an assumption that food production for the global population drives global land use. Land for bioenergy crops rapidly expands from 2030 to 2050, reaching a maximum of 550 Mha by 2060 in the 1.5 degree scenario, and declining to 430 Mha by 2100. The scenarios were run though a Dynamic Global Vegetation Model JULES (Harper et al. 2016) driven by future climate scenarios that cover the full range of climate sensitivity and spatial patterns from 45 Earth System Models (ESMs) in the CMIP5 database (Taylor et al. 2012) interpolated via the IMOGEN pattern-scaling method (Huntingford et al. 2010)**

*2.7.1.2.6 Enhanced weathering*

Weathering is the natural process of rock decomposition via chemical and physical processes in which $CO_2$ is removed from the atmosphere and converted to bicarbonates and/or carbonates (IPCC 2005). Mineral weathering can be accelerated through grinding up rock material to increase the surface area, and distributing it over land to provide carbon removals of 0.72–95 GtCO2 $yr^{-1}$ (Hartmann et al. 2018a; Beerling et al. 2018; Hartmann et al. 2013; Strefler et al. 2018; Manning 2008) (low evidence, low agreement). While the geochemical potential is quite large, agreement on the technical potential is low due to a variety of unknown parameters and of limits such as rates of mineral extraction, grinding, delivery, and challenges with scaling and deployment.

## 2.7.2 Integrated pathways for climate change mitigation

Land-based response options have the potential to interact, resulting in additive effects (e.g., climate co-benefits) or negating each other (e.g., through competition for land), they also interact with mitigation options in other sectors (such as energy or transport), thus they need to be assessed collectively under different climate mitigation targets and in combination with other sustainability goals (Popp et al. 2017;

Obersteiner et al. 2016; Humpenöder et al. 2018). Integrated Assessment Models (IAMs) with distinctive land-use modules are the basis for the assessment of mitigation pathways as they combine insights from various disciplines in a single framework and cover the largest sources of anthropogenic GHG emissions from different sectors (see also SR1.5 Chapter 2 and technical Annex for more details). IAMs consider a limited, but expanding, portfolio of land-based mitigation options. Furthermore, the inclusion and detail of a specific mitigation measure differs across IAMs and studies (see also SR1.5 and Chapter 6). For example, the IAM scenarios based on the Shared Socio-economic Pathways (SSPs) (Riahi et al. 2017) include possible trends in agriculture and land use for five different socioeconomic futures, but cover a limited set of land-based mitigation options: dietary changes, higher efficiency in food processing (especially in livestock production systems), reduction of food waste, increasing agricultural productivity, methane reductions in rice paddies, livestock and grazing management for reduced methane emissions from enteric fermentation, manure management, improvement of N-efficiency, 1st generation of biofuels, avoided deforestation, afforestation, bioenergy and BECCS (Popp et al. 2017). However, many "natural climate solutions" (Griscom et al. 2017), such as soil carbon management or wetland management, are not included in these scenarios.

Mitigation pathways, based on IAMs, are typically designed to find the least cost pathway to achieve a pre-defined climate target (Riahi et al. 2017). Such cost-optimal mitigation pathways, especially in RCP2.6 (broadly a 2˚ target) and 1.9 scenarios (broadly a 1.5˚ target), project GHG emissions to peak early in the 21st century, strict GHG emission reduction afterwards and, depending on the climate target, net carbon dioxide removal (CDR) from the atmosphere in the second half of the century (see Chapter 2 of SR1.5, (Tavoni et al. 2015; Riahi et al. 2017). In most of these pathways, land use is of great importance because of its mitigation potential (see Figure 2.36).: large-scale afforestation and reforestation removes substantial amounts of $CO_2$ from the atmosphere; biomass grown on cropland or from forestry residues can be used for energy generation or BECCS substituting fossil fuel emissions and generating CDR; non-$CO_2$ emissions from agricultural production can be reduced, even under improved agricultural management (Popp et al. 2017; Rogelj et al. 2018a; Van Vuuren et al. 2018).

From the IAM scenarios available to this assessment, a set of feasible mitigation pathways has been identified which is illustrative of the range of possible consequences on land use and GHG emissions (presented in this chapter) and sustainable development (see Chapter 6). An assessment based on IAM SSPs (Popp et al. 2017; Rogelj et al. 2018b) highlighted the importance of socio-economic baseline conditions, international cooperation, timing and sectoral participation for climate change mitigation as well as specifications of climate long-term goals land-related GHG emissions and land dynamics (Popp et al. 2017; Rogelj et al. 2018a). Thus, the IAM scenarios selected here vary due to underlying socio-economic and policy assumptions, mitigation options considered, long-term climate goal, the level of inclusion of other sustainability goals (such as land and water restrictions for biodiversity conservation or food production), and models by which they are generated.

In the baseline case without climate change mitigation, global $CO_2$ emissions from land-use change decrease over time – turning negative by the end of the century. Median global $CO_2$ emissions from land-use change across 5 SSPs and 5 IAMs decrease from 3 to1.9 to -0.7 Gt $CO_2$-eq $yr^{-1}$ in 2030, 2050 and 2100 respectively (Figure 2.34). In contrast, $CH_4$ and $N_2O$ emissions from agricultural production remain rather constant over time ($CH_4$: 6, 6.5 and 5.9 Gt $CO_2$-eq $yr^{-1}$ in 2030, 2050 and 2100 respectively; $N_2O$: 2.4, 2.7 and 2.7 Gt $CO_2$ -eq $yr^{-1}$ in 2030, 2050 and 2100 respectively).

In the mitigation cases (RCP4.5, RCP2.6 and RCP1.9), most of the scenarios indicate strong reductions in $CO_2$ emissions due to i) avoided deforestation and ii) carbon uptake due to afforestation. However, $CO_2$ emissions from land use can occur in some mitigation scenarios as a result of weak land-use change regulation (Fujimori et al. 2017, Calvin et al. 2017) or displacement effects into pasture land caused by high bioenergy production combined with forest protection only (Popp et al. 2014). The level of carbon dioxide removal globally increases with the stringency of the climate target for both afforestation (-1.3, -1.7 and -2.4

Gt $CO_2$-eq $yr^{-1}$ in 2100 for RCP4.5, RCP2.6 and RCP1.9 respectively) and BECCS (6.5, 11 and 15.3 Gt $CO_2$ $yr^{-1}$ sequestered in 2100 for RCP4.5, RCP2.6 and RCP1.9 respectively). In the mitigation cases, $CH_4$ and $N_2O$ emissions are remarkably lower compared to the baseline cases ($CH_4$: 3.7, 3 and 2.1 Gt $CO_2$-eq $yr^{-1}$ in 2100 for RCP4.5, RCP2.6 and RCP1.9 respectively; $N_2O$: 2, 1.6 and 1.2 Gt $CO_2$-eq $yr^{-1}$ in 2100 for RCP4.5, RCP2.6 and RCP1.9 respectively). The reductions in the mitigation cases are mainly due to improved agricultural management such as improved nitrogen fertiliser management, improved water management in rice production, improved manure management by for example covering of storages or adoption of biogas plants, better herd management and better quality of livestock through breeding and improved feeding practices. In addition, dietary shifts away from emission-intensive livestock products also lead to decreased $CH_4$ and $N_2O$ emissions especially in RCP2.6 and RCP1.9 scenarios. However, high levels of bioenergy production can result in increased $N_2O$ emissions due to N fertilisation of dedicated bioenergy crops.

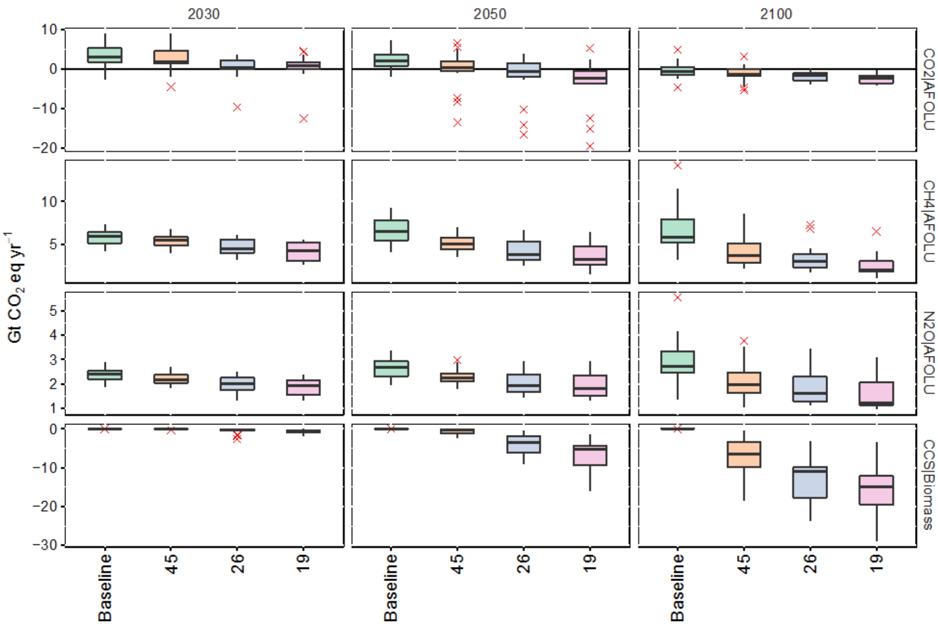


**Figure 2.34 Land-based GHG emissions and removals in 2030, 2050 and 2100 for Baseline, RCP4.5, RCP2.6 and RCP1.9 based on the Shared Socioeconomic Pathways (SSP)** (Popp et al. 2017; Rogelj et al. 2018a; Riahi et al. 2017)**. Boxplots (Tukey style) show median (horizontal line), interquartile range IQR (box) and the range of values within 1.5 x IQR at either end of the box (vertical lines) across 5 SSPs and across 5 IAMs. Outliers (red crosses) are values greater than 1.5 x IQR at either end of the box. $CH_4$ and $N_2O$ emissions are converted to $CO_2$-eq using GWP factors of 28 and 265 respectively**

Such high levels of carbon dioxide removal through mitigation options that require land conversion (BECCS and afforestation) can shape the land system dramatically (Figure 2.35). In the different SSPs and across different RCPs, the median change in global forest area ranges from -204 Mha up to +718 Mha in 2100 compared to 2010, and agricultural land used for 2nd generation bioenergy crop production ranges from 136 to 665 Mha by 2100 (Popp et al. 2017; Rogelj et al. 2018a). Land requirements for bioenergy and afforestation for a RCP1.9 scenario are higher than for a RCP2.6 and especially a RCP4.5 mitigation scenario. As a consequence of the expansion of mainly land-demanding mitigation options, global pasture land is reduced in most mitigation scenarios much stronger compared to baseline scenarios (median reduction of 4, 256, 514 and 746 Mha between 2010 and 2100 in Baseline, RCP4.5, RCP2.6 and RCP1.9 respectively). In addition, cropland for food and feed production decreases with the stringency of the climate target (+123, +21, -183, -400 Mha in 2100 compared to 2010 in Baseline, RCP4.5, RCP2.6 and RCP1.9 respectively). These reductions in agricultural land for food and feed production are facilitated by agricultural intensification on agricultural land and in livestock production systems (Popp et al. 2017) but also by changes in consumption patterns (Fujimori et al. 2017; Frank et al. 2017b). The pace of projected land-use change over the coming decades in ambitious mitigation scenarios goes well beyond historical changes in some instances (Turner et al. 2018), see also SR1.5). This raises issues for societal acceptance, and distinct policy and governance for avoiding negative consequences for other sustainability goals (Humpenöder et al. 2018; Obersteiner et al. 2016; Calvin et al. 2014), see Chapter 6 and 7).

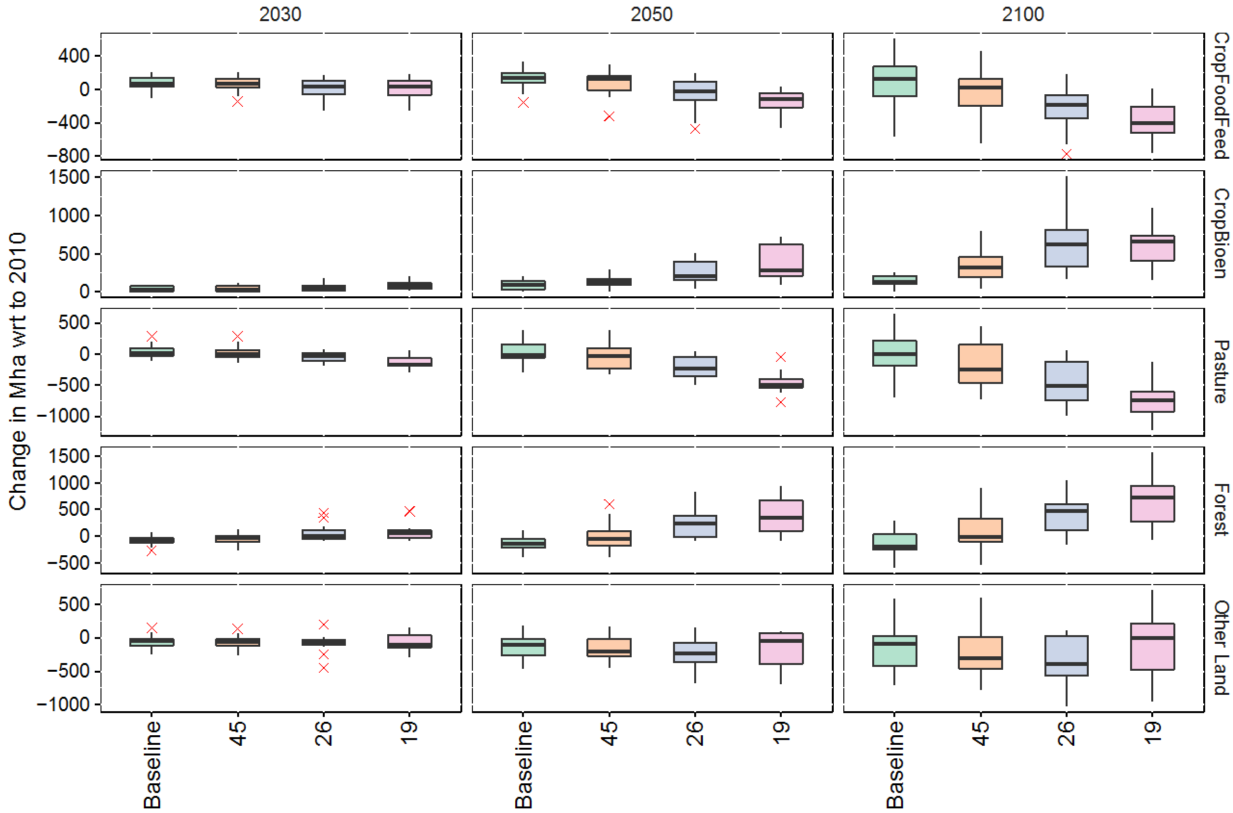


**Figure 2.35 Land-use change in 2030, 2050 and 2100 relative to 2010 for Baseline, RCP4.5, RCP2.6 and RCP1.9 based on the Shared Socioeconomic Pathways (SSP) (Popp et al. 2017; Rogelj et al. 2018a; Riahi et al. 2017). Boxplots (Tukey style) show median (horizontal line), interquartile range IQR (box) and the range of values within 1.5 x IQR at either end of the box (vertical lines) across 5 SSPs and across 5 IAMs. Outliers (red crosses) are values greater than 1.5 x IQR at either end of the box. In 2010, pasture was estimated to cover about 3-3.5 $10^3$ Mha, food and feed crops about 1.5-1.6 $10^3$ Mha, energy crops about 0-14 Mha and forest about 3.7-4.2 $10^3$ Mha, across the IAMs that reported SSP pathways (Popp et al. 2017)**

Figure 2.36 shows six alternative pathways (archetypes) for achieving ambitious climate targets highlighting AFOLU strategies and GHG emission. All pathways are assessed by different models but are all based on the Shared Socioeconomic Pathway 2 (SSP2) (Riahi et al. 2017), with all based on an RCP) 1.9 mitigation pathway expect for Pathway 1, which is RCP2.6. All scenarios use Carbon Dioxide Removal (CDR), but the amount varies across pathways, as do the relative contributions of different land-based CDR options.

Pathway1 RCP2.6 "portfolio" shows steady decrease in $CO_2$ emissions from land-use change due to measures on avoiding deforestation as well as slightly decreasing $N_2O$ and $CH_4$ emissions from agricultural production due to improved agricultural management as well as dietary shifts away from emissions-intensive livestock products. However, in contrast to $CO_2$ emissions, which turn net negative in 2050, $CH_4$ and $N_2O$ emissions remain throughout the century due to difficulties of eliminating these residual emissions based on existing agricultural management methods (Stevanović et al. 2017; Frank et al. 2017b). In addition to abating GHG emissions as well as increasing the terrestrial sink, this example also shows the importance of the land sector in providing biomass for BECCS and hence CDR in the energy sector. In this scenario, BECCS-based $CO_2$ sequestration is about 2-times higher than afforestation-based $CO_2$ sequestration in 2100. Based on these GHG dynamics, the AFOLU sector turns GHG neutral in 2100. However, accounting also for BECCS $CO_2$ withdrawal based on biomass provided by the land sector turns the land sector neutral already in 2060 significantly net-negative by the end of the century.

Pathway 2, RCP1.9 “increasing ambition” (Rogelj et al. 2018a) has dynamics of AFOLU emissions that are very similar to those in Pathway1 (RCP2.6) but all GHG emission reductions as well as terrestrial and BECCS carbon sequestration start earlier in time at a higher rate of deployment. For Pathway3, RCP 1.9 “only BECCS” in contrast to Pathway 2, CDR is only based on BECCS (Kriegler et al. 2017). In consequence, $CO_2$ emissions are persistent much longer, predominantly from indirect land-use change due to large-scale bioenergy cropland expansion into non-protected natural areas (Popp et al. 2017; Calvin et al. 2014). Nevertheless, CDR rates in 2100 are similar to Pathway 2.

Pathway 4 RCP1.9 “Early CDR” (Bertram et al. 2018) indicates that a significant reduction in the later century in the BECCS-related CDR as well as CDR in general can be achieved with earlier and mainly terrestrial CDR, starting already in 2030. In this scenario, terrestrial CDR is based on afforestation but could also supported by soil organic carbon sequestration (Paustian et al. 2016) or other natural climate solutions (Griscom et al. 2017). This scenario highlights the importance of the timing for CDR based mitigation pathways (Obersteiner et al. 2016).

In Pathway 5 RCP1.9 “low residual emissions” (Van Vuuren et al. 2018), land-based mitigation is driven by stringent enforcement of measures and technologies to reduce end-of-pipe non-$CO_2$ emissions and by introduction of in vitro (cultured) meat, reducing residual $N_2O$ and $CH_4$ emissions from agricultural production. In consequence, much lower amounts of CDR from afforestation and BECCS are needed with much later entry points to compensate for residual emissions. Finally, Pathway6 RCP1.9 “Low Energy” (Grubler et al. 2018) indicates the importance of other sectoral GHG emission reductions for the AFOLU sector. In this example, rapid reductions in energy demand and associated drops in energy related $CO_2$ emissions decrease the requirements for land-demanding CDR such as biomass production for BECCS and afforestation.

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

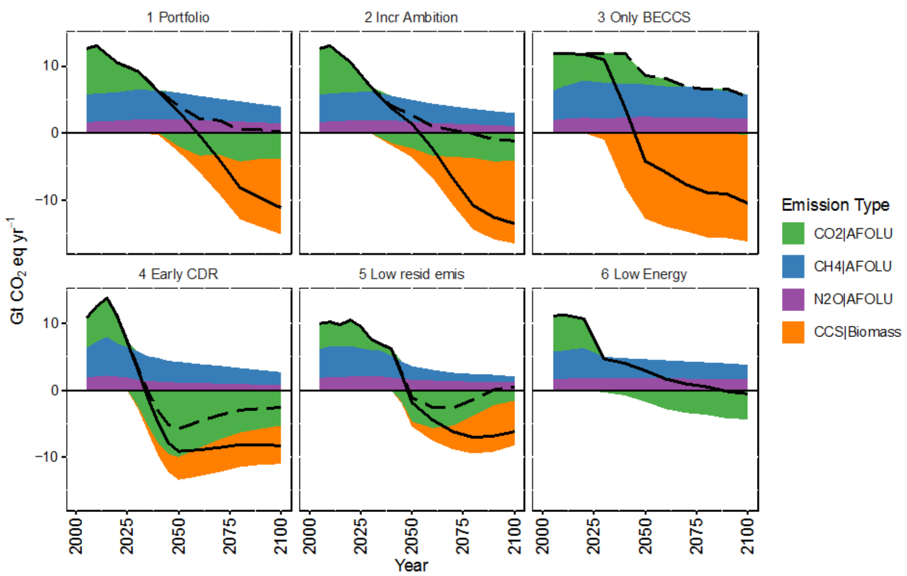

**Figure 2.36 Evolution and break down of global AFOLU GHG emissions under six alternative mitigation pathways, which illustrate the differences in timing and magnitude of AFOLU mitigation approaches including BECCS. All pathways are based on different IAM realisations of SSP2. Pathway 1 is based on RCP 2.6, while all other pathways are based on RCP 1.9. Pathway 1: MESSAGE-GLOBIOM (Fricko et al. 2017); Pathway 2: MESSAGE-GLOBIOM** (Rogelj et al. 2018a)**; Pathway 3: REMIND-MAgPIE (Kriegler et al. 2017); Pathway 4: REMIND-MAgPIE (Bertram et al. 2018); Pathway 5: IMAGE (Van Vuuren et al. 2018); Pathway 6: MESSAGE-GLOBIOM (Grubler et al. 2018). Solid lines show the net effect across AFOLU emissions and CCS|Biomass, while dashed lines show the net effect of AFOLU emissions only**

Land sector mitigation is central to the Paris Agreement temperature goals and to meeting "…*a balance between anthropogenic emissions by sources and removals by sinks of greenhouse gases in the second half of this century* …" (Article 4)(Wigley 2018). Since AFOLU is responsible for a quarter or GHG emission and the land is a sink for around a quarter of anthropogenic $CO_2$ emissions (Section 2.4), it is not surprising that land-based response options already feature prominently in the pledges countries have made under the Paris Agreement – their Nationally Determined Contributions (NDCs).

## 2.7.3 The contribution of land-based mitigation options to the Paris Agreement

Land sector mitigation is central to the Paris Agreement temperature goals and to meeting "…*a balance between anthropogenic emissions by sources and removals by sinks of greenhouse gases in the second half of this century* …" (Article 4). The balance could be interpreted in many ways e.g., does "anthropogenic" apply to both sources and sinks, how can anthropogenic fluxes be separated from natural fluxes (see 2.4.2), which GHGs are included, should non GHG climate forcers included, how is it balanced across different GHGs (metrics of $CO_2$ equivalence), is the balance to be achived at global or national levels, what time scales need to be considered (from short-term forcings to long-term permanence of storage systems), is overshooting temperature targets allowed (Fuglestvedt et al. 2018; Tanaka and O'Neill 2018; Wigley 2018; Mengel et al. 2018). Such issues can greatly affect the nature and timing of the balance, thus robust and transparent definitions of system boundaries and of reporting methods, and strong international coordination will be necessary to acheive the balance, however defined (Fuglestvedt et al. 2018).

Assuming anthropogenic applies to both emissions and removals, and natural sinks are excluded, the fact that some sector emissions are hard to eliminate entirely (e.g., air transport, food production), implies the ongoing need for anthropogenic sinks as well as reduced emissions.

### *2.7.3.1 Land mitigation in countries' Nationally Determined Contributions (NDCs)*

The land sector is expected to deliver between 20 and 25% of mitigation pledged in the NDCs based on early assessments of "Intended" NDCs, submitted ahead of the Paris Agreement and updates immediately after *(high agreement, medium certainty)* (Grassi et al. 2017; Forsell et al. 2016). Most NDCs focused on the role of forests in LULUCF, few included soil carbon sequestration or agricultural mitigation. None mention bioenergy, but this could be implicitly included with reduced emissions in energy sector through fuel substitution. 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. The countries indicating most LULUCF mitigation 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. 2012).

**Figure 2.37** shows the mitigation potential of NDCs compared to LULUCF flux based on current policies and to country-stated Business As Usual (BAU) activities (mostly not including current mitigation policies) (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 ± 0.68 $GtCO_2$-eq $yr^{-1}$, which rises to -1.14 ± 0.48 $GtCO_2e$ $yr^{-1}$ in 2030 under additional "conditional" activities. This compares to net LULUCF in 2010 calculated from the GHG Inventories of 0.01 ± 0.86 $GtCO_2$-eq $yr^{-1}$ (Grassi et al. 2018) (Figure 2.8). Forsell et al. (2016) similarly find a reduction in 2030 compared to 2010 of 0.5 $GtCO_2$-eq $yr^{-1}$ (range: 0.2–0.8) by 2020 and 0.9 Gt $CO_2$e $yr^{-1}$ (range: 0.5–1.3) by 2030 for unconditional and conditional cases.

The approach to calculating the LULUCF towards the NDC target by countries can result in a threefold difference in estimated mitigation in 2030 (1.2–3.8 $GtCO_2$-eq $yr^{-1}$), with implications for transparency (Grassi et al. 2017).

- 1.2–1.9 $GtCO_2$-eq $yr^{-1}$ in 2030 compared to 2005 emissions
- 0.7–1.4 $GtCO_2$-eq $yr^{-1}$ compared to "current activity" or "pre-INDC" reference scenario
- 2.3–3.0 $GtCO_2$-eq $yr^{-1}$ compared to country stated "BAU" reference scenario
- 3.0–3.8 $GtCO_2$-eq $yr^{-1}$ based on the countries' approach to calculating LULUCF contributions

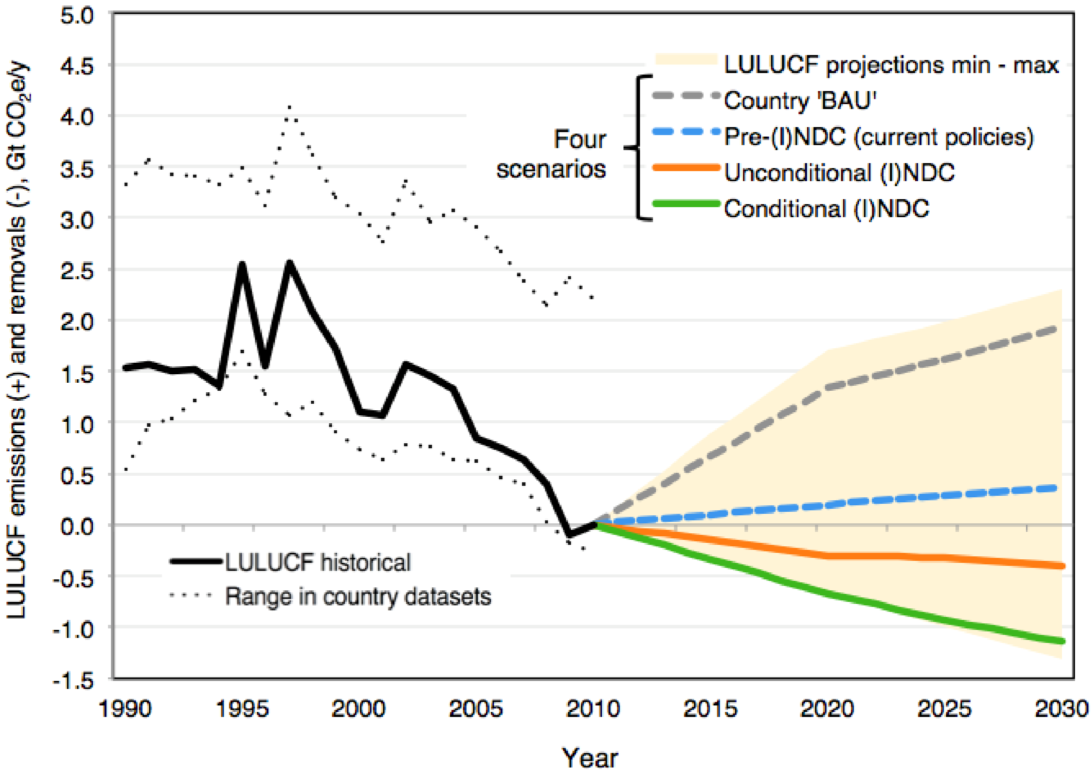


**Figure 2.37 Global Land Use, Land Use Change and Forestry (LULUCF) net greenhouse gas flux for the historical period and future scenarios based on analyses of countries' documents and mitigation pledges (Intended) Nationally Determined Contributions ((I)NDCs). The LULUCF historical data (black solid line) reflect the following countries' documents (in order of priority): data submitted to UNFCCC ((I)NDCs[2], 2015 GHG Inventories[3], recent National Communications[4],[5]); 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 2015) . 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**

[2] FOOTNOTE: UNFCCC. INDCs as communicated by Parties, http://www4.unfccc.int/submissions/indc/Submission%20Pages/submissions.aspx. (UNFCCC, 2015).
[3] FOOTNOTE: UNFCCC. Greenhouse Gas Inventories, http://unfccc.int/national_reports/annex_i_ghg_inventories/national_inventories_submissions/items/8812.php. (UNFCCC, 2015).
[4] FOOTNOTE : UNFCCC. National Communications Non-Annex 1, http://unfccc.int/nationalreports/non-annexinatcom/submittednatcom/items/653.php (UNFCCC, 2015).
[5] FOOTNOTE : UNFCCC. National Communications Annex 1, http://unfccc.int/nationalreports/annexinatcom/submittednatcom/items/7742.php; (UNFCCC, 2015).

**Update Reports[6] and National Communications), and show: the BAU scenario as defined by the country (country BAU); the trend based on pre-(I)NDC levels of activity (current policies in place in countries); and the unconditional (I)NDC and conditional (I)NDC scenarios - assuming that all countries implement the targets. 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**

#### *2.7.3.2 Raising ambition in the land sector to close the emission gap:*

*[note, analysis could still be done of IAM scenarios in database for SR1.5 report – the mitigation potential analysis is already in 2.7.2. but we want to further do an analysis of the contribution of land mitigation to closing the "gap" between current efforts and Paris targets, similar to the one from UNEP GAP report and Roe et al presented below].*

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 Global Stocktake is potentially the real "engine" of the Paris Agreement, because any identified "gap" between "collective progress" and the "well-below 2°C trajectory" is expected to motivate increased mitigation ambition by countries in successive rounds of NDCs. This means issues around uncertainties in estimating land sector emissions (Section 2.4) but equally mitigation, will be key to transparency and credibility.

The submitted Nationally Determined Contributions (NDCs) across all sectors, currently fall short of what is required to meet 2 degree or 1.5 degree pathways *(robust evidence, high agreement)* (UNEP 2017; Rogelj et al. 2016). Current commitments are more compatible with 2.5°C–3°C of warming by 2100 (Schleussner et al. 2016). To limit warming to 1.5°C (and 2°C) it would require countries to submit (and achieve) 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. 2018; Rogelj et al. 2016).

Land-based response options could provide a third of the mitigation needed **in the near term (2030 to 2050)** to close the gap between current polic**y trajectories** and what is required to achieve the Paris targets *(medium evidence, high agreement)*. The UNEP Emissions Gap Report (UNEP 2017) estimates that land-based response options offer an annual reduction potential in 2030 of agriculture 6.7 (4.9 –8.5) $GtCO_2$-eq $yr^{-1}$, forests 5.3 (4.1–6.5) $GtCO_2$-eq $yr^{-1}$, bioenergy 0.9 $GtCO_2$-eq $yr^{-1}$, and BECCS 0.3 (0.2 to 0.4) $GtCO_2$-eq $yr^{-1}$, out of a total (all sector) potential of 38 (35–41) $GtCO_2$-eq $yr^{-1}$. Thus land-based response options account for 35% of potential reduction (or 32% without bioenergy and BECCS). Roe et al. (2018) (See Figure 2.38) estimate land-based response options could provide 17 $GtCO_2$-eq $yr^{-1}$ in 2050, up 33% of total mitigation required to meet a 2˚C target. The paper details the priority regions and activity types for each response option.

Land sector response options are not a substitute for strong action in the energy and industrial sectors, as all will be needed. They rather provide near term solutions while other options are being developed and deployed.

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

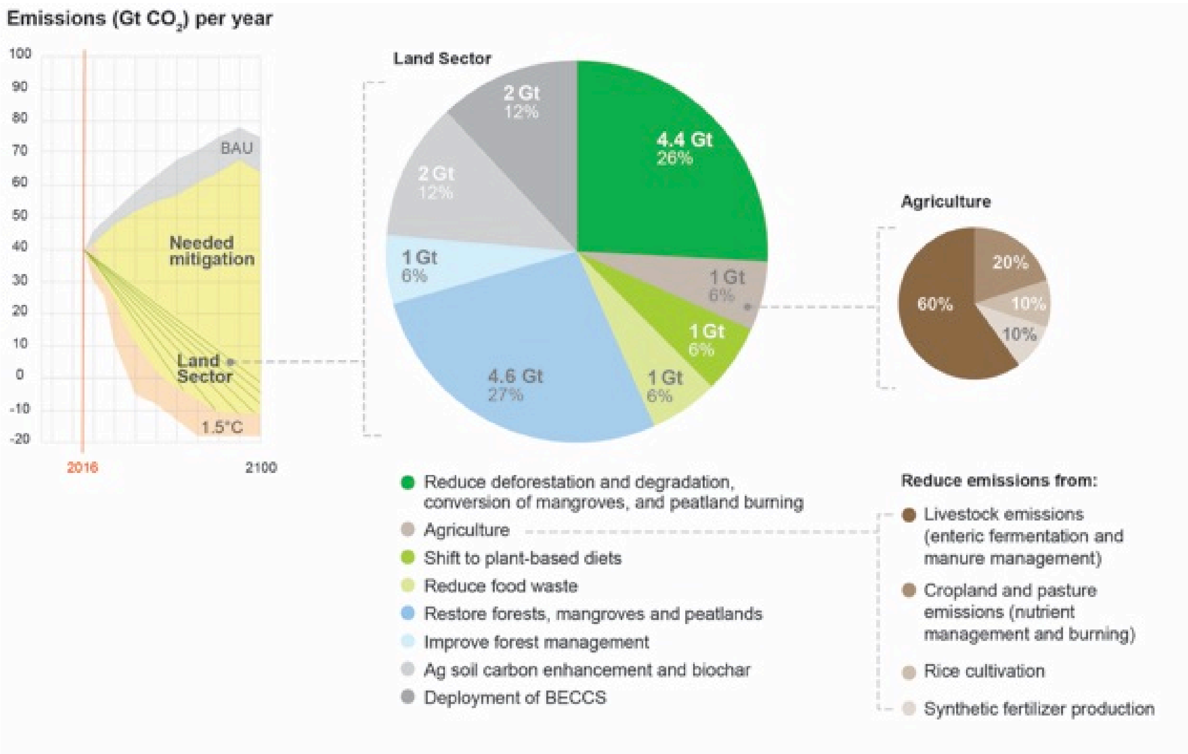


**Figure 2.38 Land-based mitigation wedges and available strategies to deliver total mitigation of about 17 $GtCO_2$-eq $yr^{-1}$ in 2050**

The wedges are measures which are individually accounted for with the intent of avoiding counting of emissions reductions. Demand-side measures only account for mitigation from overall reductions (of GHG-intensive foods and food loss and waste), and do not include efficiency or LUC mitigation.

## Box 2.1 Methodological Approaches for estimating national to global scale anthropogenic land carbon fluxes

**Bookkeeping/accounting models** (Houghton et al. 2012; 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 (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, $CO_2$, N 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 the indirect effects of climate and $CO_2$ - the "land sink". Thus, indirect effects are explicitly included. This approach also includes a "lost atmospheric sink capacity", or the carbon uptake due to environmental effects on forests that does not happen once the forests are removed (Pongratz et al., 2009).

**Integrated Assessment Models (IAMs)** use story-lines to construct alternative future scenarios of GHG emissions and atmospheric concentrations within a global socio-economic framework, including projections of AFOLU based on assumptions of, for example, crop yields, population growth, bioenergy use (See Cross-chapter Box 2: Scenarios, Chapter 1). Some models include simplified DGVMs, which may include climate and $CO_2$ effects, while others use AFOLU emissions from other sources.

**Earth System Models (ESMs)** couple (often simplified) DGVMs with a climate model, enabling exploration of feedbacks between climate change and the carbon cycle (e.g., warming effects increase respiration and lead to higher $CO_2$ concentrations, which in turn promote plant growth) (Friedlingstein et al., 2014). They sometimes include experiments run with and without land-use change to diagnose the anthropogenic AFOLU flux.

**Satellite Data** can be used as a proxy for plant activity (greenness) and to map land cover, vegetation fires and biomass density. Algorithms, models and independent data are used to calculate fluxes of $CO_2$ from satellite data, although calculating the net carbon flux is difficult because of the lack of information on the respiratory flux. Some active satellite sensors (LiDAR) are able to measure three-dimensional structure in woody vegetation, which is closely related to biomass density (Zarin et al. 2016; Baccini et al. 2012; Saatchi et al. 2011). Together with land-cover change data, these estimates of biomass density can be used to provide observational-based estimates of fluxes due to changes in forest area (e.g. (Tyukavina et al. 2015; Harris et al. 2015; Baccini et al. 2012) or degradation (Baccini et al. 2017). Satellite estimates of biomass vary considerably (Mitchard et al., 2013; Saatchi et al., 2015; Avitabile et al., 2016); data are available only for recent decades; methods generally assume that all losses of carbon are immediately released to the atmosphere; and belowground biomass and soil carbon changes have to be modelled. The approach implicitly includes indirect and natural disturbance effects as well as direct anthropogenic effects.

**Atmospheric Inversions** use observations of atmospheric concentrations with a model of atmospheric transport, based on data for wind speed and direction, to calculate the implied emissions (Gatti et al., 2014; van der Laan-Luijkx et al., 2015; Liu et al., 2017). Since AR5 there has been an increase in availability of concentration data from flux tower networks and satellites, enabling better global coverage at finer spatial scales and some national estimates (e.g. in the UK inverse techniques are used together with national GHG inventories). A combination of concentrations of different gases and isotopes enables the separation of fossil, ocean and land fluxes. However, inversions give only the net flux of $CO_2$ from land; they cannot separate natural and anthropogenic fluxes.

**Micrometeorological flux measurements**: Data on $CO_2$ concentrations and air movements recorded on instrumented towers enable calculation of $CO_2$ flux at the ecosystem scale. Global and regional Flux Networks (FluxNet (Global), AsiaFlux, Ameriflux (North America), ICOS (EU), NEON (USA), and others) contribute to a global flux data base, which is used to verify the results of modeling, inventory and remote sensing studies.

**FAOSTAT:** The United Nations Food and Agricultural Organization has produced country level estimates of greenhouse gas emissions (FAOSTAT, 2018) from agriculture (1961-2016) and land use (1990-2016) using a globally consistent methodological approach based largely on IPCC Tier 1 methods of the 2006 IPCC Guidelines (FAO, 2015; Tubiello, 2018). FAO emissions estimates were used as one of the three database inputs into the AR5 WGIII AFOLU chapter. Non-CO2 emissions from agriculture are estimated directly from national statistics of activity data reported by countries to FAO. $CO_2$ emissions from land use and land use change are computed mostly at Tier 1, albeit at fine geospatial scales to capture effects of peatland degradation and biomass fires (Rossi et al., 2016). Emissions from forest land and deforestation are based on the IPCC carbon stock change method, thus constituting a Tier 3 estimate relying of country statistics of carbon stocks and forest area collected through the FAO FRA. The carbon flux is estimated assuming instantaneous emissions in the year of forest area loss, and changes in carbon stocks within extant forests, but does not distinguish “managed” and “unmanaged” forest areas, albeit it treats separately emissons from prmary, secondary and planted forest (Federici et al. 2015).

**Country Reporting of GHG Inventories (GHGIs):** All Parties to the UNFCCC are required to report national GHG Inventories (GHGIs) of anthropogenic emissions and removals. Reporting requirements are differentiated between developed and developing countries. Because of the difficulty of separating direct anthropogenic fluxes from indirect or natural fluxes, the IPCC (2003) adopted the "managed land" concept as a proxy to facilitate GHGI reporting. All GHG fluxes on "managed land" are defined as anthropogenic, with each country applying their own definition of "managed land" (i.e. "where human interventions and practices have been applied to perform production, ecological or social functions" (IPCC 2006)). Fluxes may be determined on the basis of changes in carbon stocks (e.g., from forest inventories) or by activity data (e.g. area of land cover change management activity multiplied by emission factors or with modellef fluxes). Depending on the specific methods used, GHGIs include all direct anthropognic effects and may include the indirect anthropogenic effects of environmental change (generally sinks) and natural effects (see Section 2.4.1.2). GHG fluxes from "unmanaged land" are not reported in GHGIs because they are assumed to be non-anthropogenic. The reported estimates may then be filtered through agreed "accounting rules" - i.e. what countries actually count towards their mitigation targets (Cowie et al., 2007) ( Lee and Sanz 2017a). The accounting aims to better quantify the additional mitigation actions by, for example, factoring out the impact of natural disturbances and forest age-related dynamics (Canadell et al. 2007; Grassi et al. 2018).

# Cross-Chapter Box 3: Fire and Climate Change

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

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

While precipitation has been the major influence on wildfire regimes in pre-Industrial times, human activities became the dominant drivers since then. There was less biomass burning during the 20th century than at any time during the past two millennia as inferred from charcoal sedimentary records (Doerr and Santín 2016), though there has been an increase in the most recent decades (Marlon et al. 2016). Trends in land area burnt have varied regionally (Giglio et al. 2013). Northern Hemisphere Africa has experienced a fire decrease of 1.7 Mha $yr^{-1}$ (–1.4% $yr^{-1}$) since 2000, while Southern Hemisphere Africa saw an increase of 2.3 Mha $yr^{-1}$ (+1.8% $yr^{-1}$) during the same period. Southeast Asia witnessed a small increase of 0.2 Mha $yr^{-1}$ (+2.5% $yr^{-1}$) since 1997, while Australia experienced a sharp decrease of about 5.5 Mha $yr^{-1}$ (-10.7% $yr^{-1}$) during 2001-11, followed by an upsurge in 2011 that exceeded the annual area burned in the previous 14 years. A recent analysis using the Global Fire Emissions Database v.4 that includes small fires concluded that the net reduction in land area burnt globally during 1998–2015 was -24.3± 8.8% (-1.35 ± 0.49% $yr^{-1}$) (Andela et al. 2017). However, from the point of fire emissions it is important to consider the land cover types which have experienced changes in area burned; in this instance, most of the declines have come from grasslands, savannas and other non-forest land cover types (Andela et al. 2017). Significant increases in forest area burned (with higher fuel consumption per unit area) have been recorded as in western and boreal North America in recent times (Abatzoglou and Williams 2016; Ansmann et al. 2018). The 2017 fire season in British Columbia, Canada, was the worst ever recorded since 1950 with at least 0.9 Mha of forest burnt and smoke from these fires reaching the stratosphere over central Europe (Ansmann et al. 2018).

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

***GHG emissions from fires***

Emissions from wildfires and biomass burning are a significant source of greenhouse gases ($CO_2$, $CH_4$, $N_2O$), CO, carbonaceous aerosols, and an array of other gases including non-methane volatile organic compounds (NMVOC) (Akagi et al. 2011; van der Werf et al. 2010). The Global Fire Emissions Database V.4 (GFDB4s) has updated fire-related carbon emission estimates biome-wise, regionally and globally, using higher resolution input data gridded at 0.25˚, a new burned area dataset with small fires, improved fire emission factors (Akagi et al. 2011; Liu et al. 2014) and better fire severity characterisation of boreal forests (van der Werf et al. 2017). The estimates for the period 1997-2016 are 2.2 Pg C $yr^{-1}$, being highest in the 1997 El Nino (3.0 Pg C $yr^{-1}$) and lowest in 2013 (1.8 Pg C $yr^{-1}$). Furthermore, fire emissions during 1997–2016 were dominated by savanna (65.3%), followed by tropical forest (15.1%), boreal forest (7.4%), temperate forest (2.3%), peatland (3.7%) and agricultural waste burning (6.3%) (van der Werf et al. 2017).

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

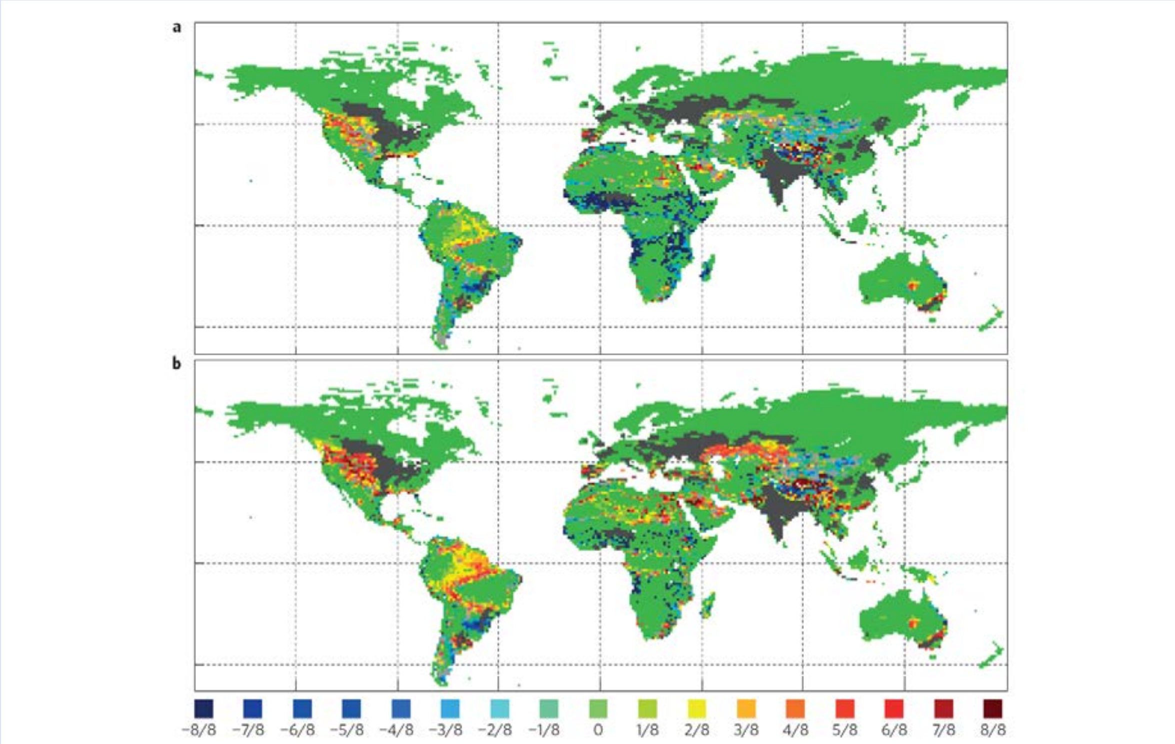
***Fires and land degradation/desertification***

Fires, especially severe ones, alter vegetation and soil properties in complex ways, both in the short- and the long-term, with consequences for carbon stock changes, albedo, fire-vegetation feedbacks and the ultimate biological capacity of the burnt land (Bond et al. 2005; Bremer and Ham 1999; MacDermott et al. 2016; Tepley et al. 2018; Moody et al. 2013; Veraverbeke et al. 2012) A fire-driven shift in vegetation from a forested state to an alternative stable state such as a grassland (Fletcher et al. 2014; Moritz 2015) with much less carbon stock is a distinct possibility. Fires also cause soil erosion through action of wind and water (Moody et al. 2013) thus resulting in land degradation (see Chapter 4) and eventually desertification (see Chapter 3).

***Fires under future climate change***

Temperature increase and precipitation decline would be the major driver of fire regimes under future climates as evapotranspiration increases and soil moisture decreases (Pechony and Shindell 2010; Aldersley et al. 2011; Abatzoglou and Williams 2016; Fernandes et al. 2017). The risk of wildfires in future could be expected to change, increasing significantly in North America, South America, central Asia, southern Europe, southern Africa, and Australia (Liu et al. 2010). There is emerging evidence that recent regional surges in wildland fires are being driven by changing weather extremes, thereby signalling geographical shifts in fire proneness (Jolly et al. 2015). Fire weather season has already increased by 18.7% globally between 1979 and 2013, with statistically significant increases across 25.3% but decreases only across 10.7% of Earth's land surface covered with vegetation; even sharper changes have been observed during the second half of this period (Jolly et al. 2015). Correspondingly, the global area experiencing long weather fire season (defined as experiencing fire weather season greater than 1 standard deviation (SD) from the mean global value) has increased by 3.1% per annum or 108.1% during 1979–2013. Fire frequencies under 2050 conditions are projected to increase by approximately 27% globally, relative to the 2000 levels, with changes in future fire meteorology playing the most important role in enhancing the future global wildfires, followed by land cover changes, lightning activities and land use, while changes in population density exhibits the opposite effects (Huang et al. 2014).

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




**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-ESM ensembles. 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)**

# Frequently Asked Questions

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

As a result of human activity, climate is changing, with impacts being seen across the globe. Land is affected by modifications in atmospheric chemistry through increased greenhouse gasses (GHGs), especially $CO_2$, and changes in climate variability, particularly extreme weather events. For example, increased concentrations of GHGs in the atmosphere has resulted in an observed "greening" of the land surface over the last 30 years due to increased $CO_2$ levels that have enhanced vegetation productivity (through photosynthesis) and water use efficiency. However, increasing GHGs have also enhanced the earth's greenhouse effect and caused the climate to warm, causing other impacts. Contemporary land cover and land use (e.g., agriculture) is adapted to current climate variability within particular temperature and/or rainfall ranges (referred to as climate envelopes) A warming climate alters the current climate variability, particularly through increases in the frequency, intensity and duration of extreme events (e.g., heat waves, very heavy rainfall, drought) and will also cause a shift of regional climate envelopes poleward and to higher elevations as these regions warm. Concurrent with these climate envelope shifts will be the emergence of new, hot climates in the tropics. These climate changes will negatively affect land use (through changes in crop productivity, irrigation needs, management practices) and land cover through loss of vegetation productivity (browning). Although $CO_2$ fertilisation has led to global greening in recent decades, increased climate variability (extremes) and climate envelope shifts as a result of the warming climate are likely to result in a greening-to-browning reversal and overwhelm any benefits to land use and land cover derived from GHG emissions.

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

Any changes to the land and how it is used can affect exchanges of water, heat, greenhouse gases (e.g., $CO_2$, $CH_4$, $N_2O$), non-greenhouse gases (e.g., biogenic volatile organic compounds – BVOCs), and aerosols (mineral, e.g., dust, or, carbonaceous, e.g., black carbon) between the land and the atmosphere. Land and land use change therefore alters the state (e.g., chemical composition, temperature and humidity) and the dynamics (e.g., strength of horizontal and vertical winds) of the atmosphere, which in turn can modify climate. Land-induced changes in heat, moisture and wind can affect neighbouring, and sometimes more distant, areas. For example, deforestation in Brazil warms the surface and enhances convection, which increases the relative temperature difference between the land and the ocean, boosting moisture advection from the ocean and thus rainfall further inland. Vegetation absorbs carbon dioxide ($CO_2$) to use for growth and maintenance. Forests contain more carbon in their biomass and soils than croplands and so a conversion of forest to cropland, for example, would result in emissions of $CO_2$ to the atmosphere. Terrestrial ecosystems are both sources and sinks of chemical compounds such as nitrogen and ozone. BVOCs contribute to forming tropospheric ozone and secondary aerosols, which can contribute to cloud formation. Semi-arid and arid regions release dust, as do cropland areas after harvest. This affects warming as both clouds and atmospheric dust affect planetary albedo (the reflectivity of the earth's surface). Although the land and land use change can affect climate, this is not a one-way interaction as changes in climate will also impact and affect the land (see FAQ 2.**1**). Understanding this two-way interaction can help improve adaptation and mitigation strategies as well as manage landscapes.

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

Renewable freshwater resources are the basis of water supply for human use (potable, agricultural and industrial) and are essential for the survival of terrestrial and aquatic ecosystems. Climate change is expected to alter the hydrological cycle and thus influence the quantity and temporal distribution of water resources, with impacts varying between regions and locations depending on dominant hydrological processes. In general, a reduction of water resources is expected in regions where rainfall is projected to be lower in the future. Additionally, a warming climate will tend to reduce surface water and groundwater resources due to an increase in evapotranspiration (and thus reduction in soil moisture and groundwater recharge), and evaporation rates from open water (rivers, lakes, wetlands) and water supply infrastructure (canals, reservoirs). Increases in the intensity of rainfall events will lead to increases in surface runoff but a reduction in soil water and groundwater recharge. In temperate mountainous locations, reduced and earlier melting snowpack will lead to summer droughts. In arid climates, higher rainfall intensities may increase groundwater resources in locations where groundwater is recharged through ephemeral streams. A warming climate will exacerbate the existing pressures on renewable freshwater resources and result in competition for water between human and natural systems.

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