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## Global Carbon and other Biogeochemical Cycles and Feedbacks

Josep G. Canadell, Pedro M.S. Monteiro, Marcos H. Costa, Leticia Cotrim Da  
Cunha, Peter M. Cox, Alexey V. Eliseev, Stephanie Henson, Masao Ishii,  
Samuel Jaccard, Charles Koven, et al.

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## AR6 WGI Report – List of corrigenda to be implemented

The corrigenda listed below will be implemented in the Chapter during copy-editing.

### CHAPTER 5

Document (Chapter, Annex, Supp. Mat...)	Section	Page :Line (based on the final pdf FGD version)	Detailed info on correction to make
Chapter 05	Table of Contents		Update Table of Contents
Chapter 5	ES	6:3-5	Replace “It is unequivocal that emissions of the well-mixed greenhouse gases (GHG) carbon dioxide (CO <sub>2</sub> ), methane (CH <sub>4</sub> ) and nitrous oxide (N <sub>2</sub> O) from human activities are the main driver of increases in atmospheric GHG concentrations since the pre-industrial period.” By “It is unequivocal that the increases in atmospheric carbon dioxide (CO <sub>2</sub> ), methane (CH <sub>4</sub> ) and nitrous oxide (N <sub>2</sub> O) since the pre-industrial period are overwhelmingly caused by human activities”
Chapter 5	ES	6:55	However, the effects of these changes is not By However, the effects of these changes are not
Chapter 5	ES	7:8	agriculture (dominated by livestock) sectors (medium confidence). By agriculture (dominated by livestock) (medium confidence).
Chapter 5	ES	9:14	a total of 655 ± 65 PgC (2390 ± 240 GtCO <sub>2</sub> ) of By a total of 655 ± 65 PgC (2390 ± 240 GtCO <sub>2</sub> ) ( <i>likely</i> range) of
Chapter 5	5.1	11 : 21	Replace “Li et al., 2016c” by “W. Li et al., 2016”
Chapter 5	5.1	11 : 22	Replace “Gruber et al., 2019a” by “Gruber et al., 2019b”
Chapter 5	5.1.1	12 : 33 12 : 38	Replace “Jones et al., 2013b” by “C.D. Jones et al., 2013”
Chapter 5	5.1.1	13 : 29	Replace “Williams et al., 2017a” by “N.L. Williams et al., 2017”
Chapter 5	5.1.2.1	15 : 17	Replace “Zhang et al., 2013” by “Y.G. Zhang et al., 2013”
Chapter 5	5.1.2.2	15 : 37- 38	Replace “Schilt et al., 2010” by “Schilt et al., 201b”
Chapter 5	5.1.2.2	16 : 31 16 : 33 17 : 16 17 : 18 17 : 19	Replace “Fischer et al., 2019b” by “H. Fischer et al., 2019”
Chapter 5	5.1.2.2	17 : 33	Replace “Li et al., 2020a” by “T. Li et al., 2020”
Chapter 5	5.1.2.2	18 : 5	Replace “Stott et al., 2019a” by “Stott et al., 2019b”
Chapter 5	5.1.2.3	18 : 49	Replace “Fischer et al., 2019b” by “H. Fischer et al., 2019”

		18 : 50-51	
Chapter 5	5.1.2.3	19 : 5	Replace “Brokin et al., 2016b” by “Brovkin et al., 2016”
Chapter 5	5.2.1.1	19:51	“all anthropogenic CO2 emissions during.” By all anthropogenic CO2 emissions.
Chapter 5	5.2.1.1	19;53	These estimates excluding By These estimates exclude
Chapter 5	5.2.1.1	20 : 5	Replace “Peters et al., 2020a” by “G.P. Peters, 2020”
Chapter 5	5.2.1.1	20 : 12	Replace “Liu et al., 2020c” by “Z. Liu et al., 2020”
Chapter 5	5.2.1.1	20 : 48	Replace “Goldewijk et al., 2017” by “Klein Goldewijk et al., 2017”
Chapter 5	5.2.1.3	23 : 36-37	Replace “Gruber et al., 2019c, 2019a” by “Gruber et al., 2019a, b”
Chapter 5	5.2.1.3	23 : 50 24 : 23-24	Replace “Resplandy et al., 2018a” by “Resplandy et al., 2018”
Chapter 5	5.2.1.3	23:46	of storage (Gruber et al., 2019a), Two definitions By of storage (Gruber et al., 2019a). Two definitions
Chapter 5	5.2.1.3	24 : 4 24 : 33	Replace “Gruber et al., 2019c” by “Gruber et al., 2019a”
Chapter 5	5.2.1.3.1	25 : 32-33 25 : 36 25 : 37 25 : 39 26 : 4-5	Replace “Gruber et al., 2019a” by “Gruber et al., 2019b”
Chapter 5	5.2.1.3.2	26 : 24-25	Replace “Gruber et al., 2019c” by “Gruber et al., 2019a”
Chapter 5	5.2.1.4.1	27 : 16-17	Replace “Le Quéré et al., 2018a” by “Le Quéré et al., 2018b”
Chapter 5	5.2.1.4.1	27 : 28-29	Replace “Palmer et al., 2019a” by “Palmer et al., 2019”
Chapter 5	5.2.1.4.1	27 : 35	Replace “Zhang et al., 2018b” by “Zhang et al., 2018”
Chapter 5	5.2.1.4.1	27 : 54	Replace “Forkel et al., 2019b” by “Forkel et al., 2019”
Chapter 5	5.2.1.4.1	27:37	observation-driven inference o By Observation-driven inference o
Chapter 5	5.2.1.4.1	28:3-4	Delete: and southeastern Australia (Canadell et al., 2020) (paper not accepted yet).
Chapter 5	5.2.1.4.1	28:30	include permafrost (Section 5.4) By include permafrost (Box 5.1)
Chapter 5	5.2.1.4.1	28 : 26-27	Replace “Pongratz et al., 2018b” by “Pongratz et al., 2018”
Chapter 5	5.2.1.4.1	28 : 27	Replace “Pugh et al., 2019a” by “Pugh et al., 2019”
Chapter 5	5.2.1.4.1	28 : 31	Replace “Koven et al., 2015a” by “Koven et al., 2015b”

Chapter 5	5.2.1.4.2	29 : 10	Replace “Liu et al., 2017” by “J. Liu et al., 2017”
Chapter 5	5.2.1.4.2	29 : 19	Replace “Cox et al, 2013a” by “Cox et al., 2013”
Chapter 5	5.2.1.4.2	29 : 20	Replace “Jung et al., 2017a” by “Jung et al., 2017”
Chapter 5	5.2.1.4.2	29 : 25	Replace “Wang et al., 2014” by “X. Wang et al., 2014”
Chapter 5	5.2.1.4.2	29:8	confirm this statement, is it “confirms this statement,” instead? please double check, it refers to “A set”
Chapter 5	Cross-Chapter Box 5.1	30:21	at the leaf, canopy to ecosystem scales By at the leaf, canopy and ecosystem scales
Chapter 5	Cross-Chapter Box 5.1	30:34	ecosystems-scale water savings By ecosystem-scale water savings
Chapter 5	CCB 5.1	30 : 37 30 : 54	Replace “Liu et al., 2020b” by “L. Liu et al., 2020”
Chapter 5	CCB 5.1	30 : 42	Replace “Y. Yang et al., 2019c” by “Y. Yang et al., 2019”
Chapter 5	CCB 5.1	31 : 21	Replace “Liu et al., 2017a” by “J. Liu et al., 2017”
Chapter 5	CCB 5.1	31 : 21	Replace “Palmer et al., 2019a” by “Palmer et al., 2019”
Chapter 5	CCB 5.1	31 : 22-23	Replace “Peters et al., 2020b” by “W. Peters et al., 2020”
Chapter 5	CCB 5.1	31 : 28	Replace “Liu et al., 2020b” by “L. Liu et al., 2020”
Chapter 5	CCB 5.1	31 : 36	Replace “Li et al., 2020b” by “X. Li et al., 2020”
Chapter 5	Cross-Chapter Box 5.1	31:22	the mid-latitudes also lead to decreased GPP By mid-latitudes also decrease GPP
Chapter 5	Cross-Chapter Box 5.1	31:23	Droughts cannot be compensated by By Droughts are not compensated by
Chapter 5	Cross-Chapter Box 5.1	32:5	electricity which is By Electricity, which is
Chapter 5	Cross-Chapter Box 5.1	32:22	precipitation can in some regions even By precipitation can, in some regions, even
Chapter 5	5.2.1.5	32:46	hemisphere By hemispheres
Chapter 5	5.2.1.5	32:50-51	from Forestry and other Land Use By From land-use, land-use change, and forestry
Chapter 5	5.2.1.5	33 : 18	Replace “Resplandy et al., 2018b” by “Resplandy et al., 2018”
Chapter 5	5.2.1.5	33 : 48	Replace “BGR (2019)” by “BGR (2020)”
Chapter 5	5.2.2	34 : 28	Replace “Ferretti et al., 2005” by “Ferreti, 2005”
Chapter 5	5.2.2	34:18	lifetime is estimated to be $9.1 \pm 0.9$ By lifetime is estimated to be $9.1 \pm 0.9$ years

Chapter 05	5.2.2	34:18	Replace "9.1 ±0.9" by "9.1 ± 0.9 years"
Chapter 05	5.2.2	34:18	Replace "(Chapter 6, Section 3.1)" by "(Chapter 6, Section 6.3.1, Table 6.2)"
Chapter 5	5.2.2.1	35:9	the scattering aerosols By the scattering of aerosols
Chapter 5	5.2.2.1	35 : 4 35 : 19	Replace "Ferretti et al., 2005" by "Ferreti, 2005"
Chapter 5	5.2.2.1	35:22-24	Replace exploitation since the early 2000s (Patra et al., 2016; Jackson et al., 2020; Chandra et al., 2021). A comprehensive assessment of the CH 4 growth rates over the past 4 decades is presented in the Cross-Chapter Box 5.2. By exploitation since the early 2000s (Patra et al., 2016; Jackson et al., 2020; Chandra et al., 2021). <b>Atmospheric concentrations of CH4 reached 1866.3 ppb in 2019 (Fig. 5.14).</b> A comprehensive assessment of the CH4 growth rates over the past <b>four</b> decades is presented in the Cross-Chapter Box 5.2.
Chapter 5		36:3	Replace "Furthermore, top-down estimates suggest emissions from China's coal mines have continued to grow at a slower rate after 2010 (Miller et al., 2019; Chandra et al., 2021)." With "Inventory-based estimates suggest that CH4 emissions from coal mining increased by 17 Tg yr <sup>-1</sup> between the periods 2002–2006 and 2008–2012 with a dominant contribution from China (Peng et al., 2016; Crippa et al., 2020; Höglund-Isaksson et al., 2020). Recent country statistics and detailed inventory-based estimate show that CH4 emissions from coal mining in China declined between 2012 and 2016 (Sheng et al., 2019; Gao et al., 2020), while atmospheric-based estimates suggest a continuation of CH4 emissions growth but at a slower rate to the year 2015 (Miller et al., 2019) and 2016 (Chandra et al., 2021)."
Chapter 5	Table 5.2	36:38	for the two recent decades from bottom–up and By for the two most recent decades for which data is available, from bottom–up and
Chapter 5	5.2.2.3	37:13	processes, that are By processes that are
Chapter 5	5.2.2.3	37:17	The wetland emissions By Wetland emissions
Chapter 5	5.2.2.3	37:21	during the La Niña By during La Niña
Chapter 5	5.2.2.4	38:40	gap between bottom-up estimations and that are used in top-down models By gap between bottom-up and top-down estimates
Chapter 5	5.2.2.4	38 : 52- 53	Replace "Zhang et al., 2020a" by "L. Zhang et al., 2020"
Chapter 5	5.2.2.4	39 : 1	Replace "Thornton et al., 2016" by "Thornton et al., 2016a"
Chapter 5	5.2.2.5	39 : 13	Replace "Wang et al., 2019b" by "X. Wang et al., 2019"
Chapter 5	5.2.2.5	39:9	estimations By estimates
Chapter 5	CCBox 5.2 Figure 1 caption	40:23	mass balance includes chemical By mass balance that includes chemical

Chapter 5	CCBox 5.2	40:38	increase in from emissions agriculture By increase in emissions from agriculture
Chapter 05	CCB 5.2	40:38	Replace "from emissions" by "emissions from"
Chapter 5	CCBox 5.2	41:1	timeseries By time series
Chapter 5	CC Box 5.2	41:37	emissions has been By emissions has
Chapter 5	5.2.3	42 : 1	Replace "Janssens-Maenhout et al., 2017" by "Janssens-Maenhout et al., 2019"
Chapter 5	5.2.3	42 : 2-3	Replace "Buitenhuis et al., 2018a" by "Buitenhuis et al., 2018"
Chapter 5	5.2.3	42 : 12	Replace "Wang et al., 2014a" by "W. Wang et al., 2014"
Chapter 5	5.2.3.1	42 : 28	Replace "Yang et al., 2020c" by "S. Yang et al., 2020"
Chapter 5	5.2.3.2	43 : 18 43 : 45	Replace "Janssens-Maenhout et al., 2017" by "Janssens-Maenhout et al., 2019"
Chapter 5	5.2.3.2	43 : 31	Replace "Buitenhuis et al., 2018b" by "Buitenhuis et al., 2018"
Chapter 5	5.2.3.3	44 : 5	Replace "Buitenhuis et al., 2018a" by "Buitenhuis et al., 2018"
Chapter 5	5.2.3.3	44 : 7	Replace "Yang et al., 2020c" by "S. Yang et al., 2020"
Chapter 5	5.2.3.3	44 : 19	Replace "Battaglia and Joos, 2018" by "Battaglia and Joos, 2018b"
Chapter 5	5.3	48:33	may enhance the emissions of By may enhance emissions of
Chapter 5	5.3.2.1	50:28	Ocean pH timeseries are are based By Ocean pH timeseries are based
Chapter 5	5.3.2.1	50 : 32 50 : 35 50 :43 50 : 51	Replace "Wu et al., 2018a" by "H.C. Wu et al., 2018"
Chapter 5	5.3.2.1	50 : 50	Replace "Pelejero, 2005" by "Pelejero et al, 2005"
Chapter 5	5.3.2.2	51 : 47	Replace "Feely et al., 2016a" by "Feely et al., 2016"
Chapter 5	5.3.2.2	52 : 7	Replace "Bates et al., 2014b" by "Bates et al. (2014)"
Chapter 5	5.3.3.1	53 : 5	Replace "Carter et al., 2017" by "Carter et al., 2017b"
Chapter 05	5.3.3.2	53:53 (duplication problem)	Replace: "They also exhibit enhanced ocean warming as a result of an increase in the equilibrium climate sensitivity (ECS) of CMIP6 relative to CMIP5 models, which contributes to increased stratification and reduced subsurface ventilation (4.3.1, 4.3.4, 5.3.3.2, 7.4.2, 7.5.6, 9.2.1, TS2.4). Consequently, CMIP6 model ensembles not only reproduce the ocean deoxygenation trend of -0.30 to -1.52 mmol m <sup>-3</sup> per decade between 1970–2010 reported in SROCC (Section 5.2.2.4) with a very likely range, but also project 32–71 % greater subsurface (100–600 m) oxygen decline relative to their RCP analogues in CMIP5, reaching to the likely range of decline of 6.4 ± 2.9 mmol m <sup>-3</sup> under SSP1–2.6 and 13.3 ± 5.3 mmol m <sup>-3</sup> under SSP5–8.5, from 1870–1899 to 2080–2099. However, they also exhibit

			<p>enhanced surface ocean warming as a result of an increased climate sensitivity (ECS), which contributes to greater reduction in subsurface ventilation. Consequently, CMIP6 model ensembles now reproduce the recent observed historical ocean deoxygenation trend of <math>-0.30</math> to <math>-1.52</math> mmol m<sup>-3</sup> per decade between 1970–2010 reported in SROCC (Section 5.2.2.4) within 90% confidence range, but project 32–71 % greater subsurface (100–600m) oxygen decline relative to their RCP analogues in CMIP5, reaching to the likely range of decline of <math>6.4 \pm 2.9</math> mmol m<sup>-3</sup> under SSP1–2.6 and <math>13.3 \pm 5.3</math> mmol m<sup>-3</sup> under SSP5–8.5, from 1870–1899 to 2080–2099 due to increased warming (Kwiatkowski et al., 2020). It is concluded that the oxygen content of subsurface ocean is projected to transition to historically unprecedented condition with decline over the 21st century (/medium confidence/).”</p> <p>By</p> <p>“They also exhibit enhanced ocean warming as a result of an increase in the equilibrium climate sensitivity (ECS) of CMIP6 relative to CMIP5 models, which contributes to increased stratification and reduced subsurface ventilation (4.3.1, 4.3.4, 5.3.3.2, 7.4.2, 7.5.6, 9.2.1, TS2.4). Consequently, CMIP6 model ensembles not only reproduce the ocean deoxygenation trend of <math>-0.30</math> to <math>-1.52</math> mmol m<sup>-3</sup> per decade between 1970–2010 reported in SROCC (Section 5.2.2.4) with a very likely range, but also project 32–71 % greater subsurface (100–600 m) oxygen decline relative to their RCP analogues in CMIP5, reaching to the likely range of decline of <math>6.4 \pm 2.9</math> mmol m<sup>-3</sup> under SSP1–2.6 and <math>13.3 \pm 5.3</math> mmol m<sup>-3</sup> under SSP5–8.5, from 1870–1899 to 2080–2099 (Kwiatkowski et al., 2020). It is concluded that the oxygen content of subsurface ocean is projected to transition to historically unprecedented condition with decline over the 21st century (/medium confidence/).”</p>
Chapter 5	5.3.3.2	54 : 36	Replace “Babbin et al., 2015a” by “Babbin et al., 2015”
Chapter 05	5.3.4.1	55:36	Citation Yamamoto et al., 2012 <a href="https://bq.copernicus.org/articles/9/2365/2012/">https://bq.copernicus.org/articles/9/2365/2012/</a> Missing from reference list
Chapter 05	5.3.4.1	55:21	In “ESMs in CMIP6 s show $+73 \pm 12\%$ increase”: Delete “s” after “CMIP6”.
Chapter 5	5.3.5.1	56 : 32 56 : 40	Replace “Sun et al., 2020a” by “H. Sun et al., 2020”
Chapter 5	5.3.5.1	56: 44	Replace “Cotovicz Jr. et al., 2015a” by “Cotovicz Jr. et al., 2015”
Chapter 5	5.3.5.1	57 : 5 57 : 8-9 57 : 16 57 : 19	Replace “Limburg et al., 2020a” by “Limburg et al., 2020”
Chapter 5	5.3.5.1	57 : 19	Replace “Naqvi et al., 2010b” by “Naqvi et al., 2010”
Chapter 5	5.3.5.2	57 : 39-40	Replace “Feely et al., 2016b” by “Feely et al., 2016”
Chapter 5	5.3.5.2	58 : 11	Replace “Cotovicz Jr. et al., 2015b” by “Cotovicz Jr. et al., 2015”
Chapter 5	5.3.5.2	58 : 24	Replace “Sun et al., 2020a” by “H. Sun et al, 2020”
Chapter 5	5.3.5.2	58 : 29	Replace “Li et al., 2016b” by “M. Li et al., 2016”
Chapter 5	5.3.5.2	58 : 47-48	Replace “Limburg et al., 2020b” by “Limburg et al., 2020”
Chapter 5	5.4.1	59 : 22	Replace “Walker et al., 2019a” by “A.P. Walker et al., 2019”
Chapter 5	5.4.1	59 : 30	Replace “Koven et al., 2015” by “Koven et al., 2015b”

Chapter 5	5.4.1	59 : 49	Replace “Yang et al., 2019b” by “X. Yang et al., 2019”
Chapter 5	5.4.1	60 : 15 60 : 19	Replace “Meyerholt et al., 2020a” by “Meyerholt et al., 2020”
Chapter 5	5.4.3.2	62 : 6-7	Replace “Walker et al., 2019b” by “X.J. Walker et al., 2019”
Chapter 5	5.4.3.3	63 : 6	Replace “Koven et al., 2015a” by “Koven et al., 2015b”
Chapter 5	5.4.3.3	63 : 33	Replace “Meyerholt et al., 2020b” by “Meyerholt et al., 2020”
Chapter 5	Box 5.1	64 : 47-48	Replace “Thornton et al., 2016a” by “Thornton et al., 2016b”
Chapter 5	Box 5.1	64:43	additional methane emissions By additional CH4 emissions
Chapter 5	Box 5.1	65:6	methane fluxes By CH4 fluxes
Chapter 5	Box 5.1	65:34	methane emissions By CH4 emissions
Chapter 5	Box 5.1	65 : 9	Replace “Thornton et al., 2016b” by “Thornton et al., 2016a”
Chapter 5	Box 5.1	65 : 39	Replace “Cias et al., 2012b” by “Cias et al., 2012”
Chapter 5	Box 5.1	67 : 6	Replace “Koven et al., 2015b” by “Koven et al., 2015a”
Chapter 5	5.4.4.1	67 : 32	Replace “Gruber et al., 2019” by “Gruber et al., 2019b”
Chapter 5		67:33	sediments also play a minor role By sediments play a minor role
Chapter 5	5.4.4.2	68 : 14	Replace “Kwiatkowski et al., 2017b” by “Kwiatkowski et al., 2017”
Chapter 5	5.4.5	69 : 18	Replace “Friedlingstein et al., 2014” by “Friedlingstein et al., 2014b”
Chapter 05	5.4.5	69:46	JAMSETC to JAMSTEC
Chapter 05	5.4.5	69:46	for MPI-ESM1.2-LR the number of PFTs from 13 to 12
Chapter 05	5.4.5	69:46	for GFDL- ESM4 the # of phytoplankton groups from 2 to 3
Chapter 05	5.4.5	70 –73	Replace the following heading numbers “5.2.1.1” to “5.4.5.1” “5.4.5.1” to “5.4.5.2” “5.4.5.2” to “5.4.5.3” “5.4.5.3” to “5.4.5.4” “5.4.5.4” to “5.4.5.5”
Chapter 5	5.4.6	75 : 44	Replace “Cox et al., 2013b” by “Cox et al., 2013”
Chapter 5	5.4.7	76 : 28	Replace “Zhang et al., 2020b” by “X. Zhang et al., 2020”
Chapter 5	5.4.7	76 : 30-31	Replace “Shindell et al., 2013a; Stocker et al., 2013a” by “Shindell et al., 2013; B.D. Stocker et al., 2013”
Chapter 5	5.4.7	76 : 32	Replace “Shindell et al., 2013a” by “Shindell et al., 2013”



Chapter 5	5.4.7	76 : 34-35	Replace “Shindell et al., 2013b; Stocker et al., 2013a” by “Shindell et al., 2013; B.D. Stocker et al., 2013”
Chapter 5	5.4.7	76 : 54	Replace “Stocker et al., 2013b; Zaehle, 2013a” by “B.D. Stocker et al., 2013, Zaehle, 2013”
Chapter 5	5.4.7	77 : 1	Replace “Fischer et al., 2019b” by “H. Fischer et al., 2019”
Chapter 5	5.4.7	77 : 14	Replace “Battaglia and Joos, 2018” by “Battaglia and Joos, 2018b”
Chapter 5	5.4.8	77 : 50	Replace “Koven et al., 2015c, 2015b” by “Koven et al., 2015a, c”
Chapter 5	5.4.8	77 : 53	Replace “Shindell et al., 2013a; Stocker et al., 2013a” by “Shindell et al., 2013; B.D. Stocker et al., 2013”
Chapter 5	5.4.8	77 : 54	Replace “Stocker et al., 2013” by “B.D. Stocker et al., 2013”
Chapter 5	5.4.8	77 : 55	Replace “Battaglia and Joos, 2018” by “Battaglia and Joos, 2018b”
Chapter 5		78:50-52	It is not currently possible to carry out a full assessment of proposed abrupt changes and tipping points in the biogeochemical cycles. In this section we therefore focus instead on estimating By <del>It is not currently possible to carry out a full assessment of proposed abrupt changes and tipping points in the biogeochemical cycles.</del> In this section we focus on estimating
Chapter 5	5.4.9.1.1	80:18	Into the previous boreal area By into previous boreal areas
Chapter 5	5.4.9.1.1	80:21	vegetation C loss By vegetation carbon loss
Chapter 5		80:38	from the permafrost are assessed By from permafrost thawing are assessed
Chapter 5	5.4.9.1.1	80 : 13	Replace “Cox et al., 2013” by “Cox et al., 2013”
Chapter 5	5.4.9.1.1	80 : 23	Replace “Walker et al., 2019a” by “A.P. Walker et al., 2019”
Chapter 5	5.4.9.1.2	80:31	methane emissions By CH4 emissions
Chapter 5	5.4.9.1.2	80:35	methane By CH4
Chapter 5	5.4.9.1.2	80:36	methane By CH4
Chapter 5	5.4.9.1.3	80:44	methane By CH4
Chapter 5	5.4.9.1.3	80:47	methane By CH4
Chapter 5	5.4.9.1.3	80:52	methane By CH4
Chapter 5	5.4.1.1	82 : 36	Replace “Li et al., 2016a; Lovenduski et al., 2019b” by “H. Li et al., 2016; Lovenduski et al., 2019a”
Chapter 5	5.4.1.1	82 : 39	Replace “Lovenduski et al., 2019b” by “Lovenduski et al., 2019a”

Chapter 5	5.4.1.1	82 : 40	Replace “Li et al., 2016a” by “H. Li et al., 2016”
Chapter 5	5.4.1.1	82 : 55	Replace “Lovenduski et al., 2019a” by “Lovenduski et al., 2019b”
Chapter 5	5.4.1.1	83 : 3	Replace “Lovenduski et al., 2019b” by “Lovenduski et al., 2019a”
Chapter 5	5.5	83 : 18	Replace “Collins et al., 2013; Stocker et al., 2013” by “” by “M. Collins et al., 2013; T.F. Stocker et al., 2013”
Chapter 5	5.5	83 : 22-23	Replace “Collins et al., 2013; Stocker et al., 2013” by “M. Collins et al., 2013; T.F. Stocker et al., 2013”
Chapter 5	5.5.1.1.	83 : 40	Replace “Collins et al., 2013a; Stocker et al., 2013c” by “Collins et al., 2013; T.F. Stocker et al., 2013”
Chapter 5	5.5.1.1	83 : 53	Replace “Williams et al., 2016, 2017c” by “R.G. Williams et al., 2016, 2017b”
Chapter 5	5.4.11	83:2	and the predictability of terrestrial ecosystem drivers such as By and the drivers of terrestrial carbon flux predictability such as
Chapter 5	5.5.1.1	84 : 16	Replace “Collins et al., 2013a” by “M. Collins et al., 2013”
Chapter 5	CCB 5.3	84 : 50	Replace “Gruber et al., 2019a” by “Gruber et al., 2019b”
Chapter 5	CCBox 5.3	85:12	However, the By The
Chapter 5	5.5.1.2.1	87 : 28	Replace “Collins et al., 2013a” by “M. Collins et al., 2013”
Chapter 5	5.5.1.2.3	86 : 47	Replace “Tokarska et al., 2019b” by “Tokarska et al., 2019a”
Chapter 5	5.5.1.2.3	88 : 49	Replace Williams et al., 2017b” by “R.G. Williams et al., 2017a”
Chapter 5	5.5.1.2.3	88 : 54	Replace “Collins et al., 2013b” by “M.. Collins et al., 2013”
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Chapter 5	5.5.1.3	89 : 43-44	Replace “Williams et al., 2017c” by “R.G. Williams et al., 2017b”
Chapter 5	5.5.1.3	89 : 51-52	Replace “Williams et al., 2017b” by “R.G. Williams et al., 2017a”
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Chapter 5	Table 5.7	90	Replace “Millar et al., 2017b” by “Millar et al., 2017a”
Chapter 5	Table 5.7	90	Replace “Williams et al., 2017c” by “R.G. Williams et al., 2017b”
Chapter 5	Table 5.7	90	Replace “Smith et al., 2018a” by “C.J. Smith et al., 2018”
Chapter 5	5.5.2	91 : 36	Replace “Millar et al., 2017b” by “Millar et al., 2017a”
Chapter 5	5.5.2	91 : 41	Replace “Collins et al., 2013b” by “M. Collins et al., 2013”
Chapter 5	5.5.1.4	91:17	1.0–2.3°C per 1000 PgC. By 1.0–2.3°C per 1000 PgC (0.27°C–0.63°C per 1000 GtCO <sub>2</sub> )
Chapter 5	5.5.2	92 : 1-2	Replace “Collins et al., 2013a; Stocker et al., 2013c” by “M. Collins et al., 2013; T.F. Stocker et al., 2013”

Chapter 5	5.5.2	92 : 3	Replace “Millar et al., 2017a” by “Millar et al., 2017b”
Chapter 5	5.5.2.1	92 : 11	Replace “Collins et al., 2013; Stocker et al., 2013” by “M. Collins et al., 2013; T.F. Stocker et al., 2013”
Chapter 5	5.5.2.1	92 : 15 92 : 16-17 92 : 30	Replace “Collins et al., 2013a” by “M. Collins et al., 2013” Replace “Stocker et al., 2013c” by “T.F. Stocker et al., 2013”
Chapter 5	5.5.2.1	92 : 33	Replace “Rogelj et al., 2015a, 2016, 2018b” by “Rogelj et al., 2015a, 2016, 2018”
Chapter 5	5.5.2.1	92 : 36	Replace “Collins et al., 2013b” by “M. Collins et al., 2013”
Chapter 5	5.5.2.1	92 : 40-41	Replace “Friedlingstein et al., 2014” by “Friedlingstein et al., 2014a”
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Chapter 5	5.5.2.1	92 : 48	Replace “Millar et al., 2017a” by “Millar et al., 2017b”
Chapter 5	5.5.2.1	93 : 1-2	Replace “Collins et al., 2013” by “M. Collins et al., 2013”
Chapter 5	5.5.2.1	93 : 21-22 93 : 24	Replace “Gidden et al., 2018” by “Gidden et al., 2019”
Chapter 5	5.5.2.2	93 : 44	Replace “Collins et al., 2013b” by “M. Collins et al., 2013”
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Chapter 5	5.5.2.2.2	94 : 3 94 : 10	Replace “Tokarska et al., 2019a” by “Tokarska et al., 2019b”
Chapter 5	5.5.2.2.2	94 : 8	Replace “Collins et al., 2013a” by “M. Collins et al., 2013” Replace “Stocker et al., 2013c” by “T.F. Stocker et al., 2013”
Chapter 5	5.5.2.2.3	94 : 20	Replace “Williams et al., 2017c” by “R.G. Williams et al., 2017b”
Chapter 5	5.5.2.2.3	94 : 34	“Millar et al., 2017a” by “Millar et al., 2017b”
Chapter 5	5.5.2.2.3	94 : 34-35	Replace “Smith et al., 2018a” by “C.J. Smith et al., 2018”
Chapter 5	5.5.2.2.5	95 : 20	Replace “Burke et al., 2017” by “Burke et al., 2017b”
Chapter 5	5.5.2.3	96:10-11	655 ± 65 PgC (1-sigma range, or 2390 ± 240 GtCO <sub>2</sub> , By 655 ± 65 PgC ( <i>likely</i> range, or 2390 ± 240 GtCO <sub>2</sub> ,
Chapter 5	5.5.2.3	Table 5.8 Page: 96 Footnote: *(2)	655 ± 65 PgC (1-sigma range, or 2390 ± 240 GtCO <sub>2</sub> , By 655 ± 65 PgC ( <i>likely</i> range, or 2390 ± 240 GtCO <sub>2</sub> ,
Chapter 5	5.5.2.3	Table 5.8 Page: 96 Footnote: *(6)	very largely range (5–95%) By very likely range (5–95%)
Chapter 5	Box 5.2	97 : 10-11	Replace “Collins et al., 2013” by “M. Collins et al., 2013” Replace “IPCC, 2013” by “IPCC, 2013b” Replace “Stocker et al., 2013” by “T.F. Stocker et al., 2013”
Chapter 5	Box 5.2	97 : 14	Replace “Millar et al., 2017b” by “Millar et al., 2017a”
Chapter 5	Box 5.2	97 : 20 97 : 22 97 : 26	Replace “Millar et al., 2017a” by “Millar et al., 2017b”

Chapter 5	Box 5.2	97 : 22	Replace “Tokarska et al., 2019a” by “Tokarska et al., 2019b”
Chapter 5	Box 5.2	97 : 29	Replace “Meinshausen et al., 2011b” by “Meinshausen et al., 2011c”
Chapter 5	Box 5.2	97 : 30	Replace “Stocker et al., 2013c” by “T.F. Stocker et al., 2013”
Chapter 5	Box 5.2	97 : 34	Replace “Meinshausen et al., 2011a” by “Meinshausen et al., 2011c”
Chapter 5	Box 5.2	97 : 34	Replace “Friedlingstein et al., 2014” by “Friedlingstein et al., 2014a”
Chapter 5	Box 5.2	97 : 51	Replace “Meinshausen et al., 2011c” by “Meinshausen et al., 2011b”
Chapter 5	Box 5.2	97 : 52	Replace “Smith et al., 2018b” by “C.J. Smith et al., 2018”
Chapter 5	5.6.2	99 : 44	Replace “IPCC, 2018a” by “IPCC, 2018”
Chapter 5	Table 5.9	100	Replace “Oschlies et al., 2010” by “Oschlies et al., 2010a”
Chapter 5	Table 5.9	101	Replace “Oschlies et al., 2010c” by “Oschlies et al., 2010b”
Chapter 5	5.6.2.1.1	103 : 22	Replace “Collins et al., 2013a” by “M. Collins et al., 2013”
Chapter 5	5.6.2.2.1	107 : 3	Replace “Pugh et al., 2019b” by “Pugh et al., 2019”
Chapter 5	5.6.2.2.1	107 : 17	Replace “Boysen et al., 2017” by “Boysen et al., 2017b”
Chapter 5	5.6.2.2.1	107 : 21-22	Replace “Zhang et al., 2013” by “W. Zhang et al., 2013”
Chapter 5	5.6.2.2.1	107 : 32 107 : 52 108 : 13-14 108 : 47	Replace “Smith et al., 2018c” or “Smith et al., 20a8” by “P. Smith et al., 2018”
Chapter 5	5.6.2.2.1	107 : 45	Replace “Sun et al., 2020b” by “W. Sun et al., 2020”
Chapter 5	5.6.2.2.1	107 : 48-49	Replace “Paustian et al., 2016a” by “Paustian et al., 2016”
Chapter 5	5.6.2.2.1	107:4	more additional By additional
Chapter 5	5.6.2.2.1	107:25	in dry areas  By  in dry regions
Chapter 5	5.6.2.2.1	107:45	agroforestry, and cropland  By  agroforestry, cropland
Chapter 5	5.6.2.2.1	107:46	conversion management  By  conversion
Chapter 5	5.6.2.2.1	108 : 4	Replace “Fischer et al., 2019a” by “B.M.C Fischer et al., 2019”

Chapter 5	5.6.2.2.1	108 : 5	Replace “Fischer et al., 2019” by “B.M.C Fischer et al., 2019”
Chapter 5	5.6.2.2.1	108 : 7	Replace “Liu et al., 2017b” by “Z. Liu et al., 2017”
Chapter 5	5.6.2.2.1	108 : 9	Replace “Wang et al., 2019a; Yang et al., 2019a” by “S. Wang et al., 2019; Yang et al., 2019”
Chapter 5	5.6.2.2.1	108 : 26	Replace “Liu et al., 2020a” by “H. Liu et al., 2020”
Chapter 5	5.6.2.2.3	110 : 4	Replace “Keller et al., 2018” by “Keller et al., 2018a”
Chapter 5	5.6.2.2.3	110 : 7	Replace “Blanc-Betes et al.” by “Blanc-Betes et al., 2020”
Chapter 5	5.6.2.2.3	110 : 11	Replace “Smith et al., 2018” by “P. Smith et al., 2018”
Chapter 5	5.6.2.2.5	110 : 45	Replace “Wu et al., 2018b” by “J. Wu et al., 2018”
Chapter 5	5.6.2.2.5	110:42-43	using for example zeolite trapping, but instead of storing it CH4 would By using, for example, zeolite trapping, but instead of storing it, CH4 would
Chapter 5	5.6.3.1	111 : 54	Replace “Yang et al., 2020a” by “C.-E. Yang et al., 2020”
Chapter 5	5.6.3.1	112 : 16 112 : 22	Replace “Yang et al., 2020b” by “C.-E. Yang et al., 2020”
Chapter 5	5.6.3.2	113 : 6 113 : 15	Replace “Yang et al., 2020b” by “C.-E. Yang et al., 2020”
Chapter 5	5.6.3.2	113 : 20	Replace “Jones et al., 2013a” by “A. Jones et al., 2013”
Chapter 5	5.7	114:7	partitioning of the land and By partitioning the land and
Chapter 5	5.7	114:13	Reduce uncertainties in wetland CH4 emissions, which is largest By Reduce uncertainties of CH4 emissions from wetlands and inland waters, which are the largest
Chapter 5	References	124:1	Two new references must be added to the list that were omitted in the FGD. These are:  Sheng, J., S. Song, Y. Zhang, R. Prinn, & G. Janssens-Maenhout. 2019. Bottom-Up Estimates of Coal Mine Methane Emissions in China: A Gridded Inventory, Emission Factors, and Trends. Environmental Science & Technology Letters 6, no. 8: 473-478. <a href="https://dash.harvard.edu/bitstream/handle/1/42497758/draft_chinacoal_final.pdf?sequence=2&amp;isAllowed=y">https://dash.harvard.edu/bitstream/handle/1/42497758/draft_chinacoal_final.pdf?sequence=2&amp;isAllowed=y</a>  Junlian Gao, ChengHe Guan, Bo Zhang, China's CH4 emissions from coal mining: A review of current bottom-up inventories, Science of The Total Environment, Volume 725, 2020, 138295, ISSN 0048-9697, <a href="https://doi.org/10.1016/j.scitotenv.2020.138295">https://doi.org/10.1016/j.scitotenv.2020.138295</a> .
Chapter 5	Figure 5.1	177	replace with updated visual roadmap, as all visual roadmaps have been harmonised (to have a set with a consistent visual identity. This does not alter the content of the chapter.)
Chapter 5	Figure 5.31	210:18	Coloured areas show the Chapter 4 assessed <i>very likely</i> range of GSAT projections and thick coloured central lines the median estimate, for each respective scenario, relative to the original scenario emissions (Riahi et al., 2017; Gidden et al., 2018; Rogelj et al., 2018a). By Coloured areas show the Chapter 4 assessed <i>very likely</i> range of GSAT projections and thick coloured central lines the median estimate, for each respective scenario. These projections are expressed relative to the cumulative CO2 emissions that are available for

			emission-driven CMIP6 ScenarioMIP experiments for each respective scenario (Riahi et al., 2017; Gidden et al., 2018; Rogelj et al., 2018a).
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**Chapter 5: Global Carbon and other Biogeochemical Cycles and Feedbacks****Coordinating Lead Authors:**

Josep G. Canadell (Australia), Pedro M.S. Monteiro (South Africa)

**Lead Authors:**

Marcos H. Costa (Brazil), Leticia Cotrim da Cunha (Brazil), Peter M. Cox (United Kingdom), Alexey V. Eliseev (Russian Federation), Stephanie Henson (United Kingdom), Masao Ishii (Japan), Samuel Jaccard (Switzerland), Charles Koven (United States of America), Annalea Lohila (Finland), Prabir K. Patra (Japan/India), Shilong Piao (China), Joeri Rogelj (United Kingdom/Belgium), Stephen Syampungani (Zambia), Sönke Zaehle (Germany), Kirsten Zickfeld (Canada/Germany)

**Contributing Authors:**

Georgii A. Alexandrov (Russian Federation), Govindasamy Bala (India/ United States of America), Laurent Bopp (France), Lena Boysen (Germany), Long Cao (China), Naveen Chandra (Japan/India), Philippe Ciais (France), Sergey N. Denisov (Russian Federation), Frank J. Dentener (EU, The Netherlands), Hervé Douville (France), Amanda Fay (United States of America), Piers Forster (United Kingdom), Baylor Fox-Kemper (United States of America), Pierre Friedlingstein (United Kingdom), Weiwei Fu (United States of America /China), Sabine Fuss (Germany), Véronique Garçon (France), Bettina Gier (Germany), Nathan P. Gillett (Canada), Luke Gregor (Switzerland/South Africa), Karsten Haustein (United Kingdom /Germany), Vanessa Haverd (Australia), Jian He (United States of America /China), Helene T. Hewitt (United Kingdom), Forrest M. Hoffman (United States of America), Tatiana Ilyina (Germany), Robert Jackson (United States of America), Christopher Jones (United Kingdom), David P. Keller (Germany/ United States of America), Lester Kwiatkowski (France/ United Kingdom), Robin D. Lamboll (United Kingdom/United States of America, United Kingdom), Xin Lan (United States of America/China), Charlotte Laufkötter (Switzerland/Germany), Corinne Le Quéré (United Kingdom), Andrew Lenton (Australia), Jared Lewis (Australia/New Zealand), Spencer Liddicoat (United Kingdom), Laura Lorenzoni (United States of America/Venezuela), Nicole Lovenduski (United States of America), Andrew H. MacDougall (Canada), Sabine Mathesius (Canada/Germany), Damon H. Matthews (Canada), Malte Meinshausen (Australia/Germany), Igor I. Mokhov (Russian Federation), Vaishali Naik (United States of America), Zebedee R. J. Nicholls (Australia), Intan Suci Nurhati (Indonesia), Michael O'Sullivan (United Kingdom), Glen Peters (Norway), Julia Pongratz (Germany), Benjamin Poulter (United States of America), Jean-Baptiste Sallée (France), Marielle Saunois (France), Edward A.G. Schuur (United States of America), Sonia I. Seneviratne (Switzerland), Ann Stavert (Australia), Parvatha Suntharalingam (United Kingdom/United States of America), Kaoru Tachiiri (Japan), Jens Terhaar (Switzerland/Germany), Rona Thompson (Norway, Luxembourg/New Zealand), Hanqin Tian (United States of America), Jocelyn Turnbull (New Zealand), Sergio M. Vicente-Serrano (Spain), Xuhui Wang (China), Rik Wanninkhof (United States of America), Phil Williamson (United Kingdom),

**Review Editors:**

Victor Brovkin (Germany/Russian Federation), Richard A. Feely (United States of America)

**Chapter Scientist:**

Alice D. Lebehot (South Africa/France)

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4 K. Zickfeld, 2021, Global Carbon and other Biogeochemical Cycles and Feedbacks. In: *Climate Change*  
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7 Péan, S. Berger, N. Caud, Y. Chen, L. Goldfarb, M. I. Gomis, M. Huang, K. Leitzell, E. Lonnoy, J.B.R.  
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1	<b>Table of content</b>	
2		
3	<b>Executive Summary</b> .....	<b>6</b>
4		
5	<b>5.1 Introduction</b> .....	<b>11</b>
6	5.1.1 The Physical and Biogeochemical Processes in Carbon-Climate feedbacks .....	12
7	5.1.2 Paleo Trends and Feedbacks .....	14
8	5.1.2.1 Cenozoic Proxy CO <sub>2</sub> Record .....	14
9	5.1.2.2 Glacial-Interglacial Greenhouse Gases Records .....	15
10	5.1.2.3 Holocene Changes .....	18
11		
12	<b>5.2 Historical Trends, Variability and Budgets of CO<sub>2</sub>, CH<sub>4</sub>, and N<sub>2</sub>O</b> .....	<b>19</b>
13	5.2.1 CO <sub>2</sub> : Trends, Variability and Budget.....	19
14	5.2.1.1 Anthropogenic CO <sub>2</sub> Emissions.....	19
15	5.2.1.2 Atmosphere .....	21
16	5.2.1.3 Ocean Carbon Fluxes and Storage.....	23
17	5.2.1.3.1 Ocean Carbon Fluxes and Storage: Global Multi-Decadal Trends.....	24
18	5.2.1.3.2 Ocean Carbon Fluxes and Storage: Regional – Global Variability .....	26
19	5.2.1.4 Land CO <sub>2</sub> Fluxes: Historical and Contemporary Variability and Trends .....	27
20	5.2.1.4.1 Trend in Land-Atmosphere CO <sub>2</sub> Exchange .....	27
21	5.2.1.4.2 Interannual variability in land-atmosphere CO <sub>2</sub> exchange.....	29
22		
23	<b>Cross-Chapter Box 5.1: Interactions between the carbon and water cycles, particularly under drought</b>	
24	<b>conditions</b> .....	<b>30</b>
25		
26	5.2.1.5 CO <sub>2</sub> Budget .....	32
27	5.2.2 CH <sub>4</sub> : Trends, Variability and Budget.....	34
28	5.2.2.1 Atmosphere .....	34
29	5.2.2.2 Anthropogenic CH <sub>4</sub> emissions .....	35
30	5.2.2.3 Land Biospheric Emissions and Sinks .....	37
31	5.2.2.4 Ocean and Inland Water Emissions and Sinks .....	38
32	5.2.2.5 CH <sub>4</sub> Budget .....	39
33		
34	<b>Cross-Chapter Box 5.2: Drivers of atmospheric methane changes during 1980–2019</b> .....	<b>39</b>
35		
36	5.2.3 N <sub>2</sub> O: Trends, Variability and Budget.....	41
37	5.2.3.1 Atmosphere .....	42
38	5.2.3.2 Anthropogenic N <sub>2</sub> O Emissions.....	43
39	5.2.3.3 Emissions from Ocean, Inland Water Bodies and Estuaries .....	44
40	5.2.3.4 Emissions and Sinks in Non-Agricultural Land.....	44

1	5.2.3.5	N <sub>2</sub> O budget.....	45
2	5.2.4	The Relative Importance of CO <sub>2</sub> , CH <sub>4</sub> , and N <sub>2</sub> O.....	47
3			
4	<b>5.3</b>	<b>Ocean Acidification and Deoxygenation.....</b>	<b>48</b>
5	5.3.1	Paleoclimate Context .....	48
6	5.3.1.1	Paleocene-Eocene Thermal Maximum .....	48
7	5.3.1.2	Last Deglacial Transition .....	49
8	5.3.2	Historical Trends and Spatial Characteristics in the Upper Ocean.....	50
9	5.3.2.1	Reconstructed Centennial Ocean Acidification Trends .....	50
10	5.3.2.2	Observations of Ocean Acidification over the Recent Decades .....	51
11	5.3.3	Ocean Interior Change .....	52
12	5.3.3.1	Ocean Memory – Acidification in the Ocean Interior .....	52
13	5.3.3.2	Ocean Deoxygenation and its Implications for GHGs.....	53
14	5.3.4	Future Projections for Ocean Acidification .....	54
15	5.3.4.1	Future Projections with Earth System Models .....	54
16	5.3.4.2	Reversal of Ocean Acidification by Carbon Dioxide Removal .....	56
17	5.3.5	Coastal Ocean Acidification and Deoxygenation.....	56
18	5.3.5.1	Drivers .....	56
19	5.3.5.2	Spatial Characteristics.....	57
20			
21	<b>5.4</b>	<b>Biogeochemical Feedbacks on Climate Change .....</b>	<b>58</b>
22	5.4.1	Direct CO <sub>2</sub> Effect on Land Carbon Uptake.....	59
23	5.4.2	Direct CO <sub>2</sub> Effects on Projected Ocean Carbon Uptake.....	60
24	5.4.3	Climate Effect on Land Carbon Uptake.....	61
25	5.4.3.1	Plant Physiology.....	61
26	5.4.3.2	Fire and Other Disturbances.....	61
27	5.4.3.3	Soil Carbon .....	62
28			
29	<b>BOX 5.1:</b>	<b>Permafrost Carbon and Feedbacks to Climate.....</b>	<b>63</b>
30			
31	5.4.4	Climate Effects on Future Ocean Carbon Uptake .....	67
32	5.4.4.1	Physical Drivers of Future Ocean Carbon Uptake and Storage .....	67
33	5.4.4.2	Biological Drivers of Future Ocean Carbon Uptake.....	67
34	5.4.5	Carbon Cycle Projections in Earth System Models.....	69
35	5.2.1.1	Evaluation of the Contemporary Carbon Cycle in Concentration-Driven Runs .....	70
36	5.4.5.1	Evaluation of Historical Carbon Cycle Simulations in Concentration-Driven Runs.....	71
37	5.4.5.2	Evaluation of Latitudinal Distribution of Simulated Carbon Sinks.....	71
38	5.4.5.3	Coupled Climate-Carbon Cycle Projections .....	72

1	5.4.5.4	Linear Feedback Analysis .....	73
2	5.4.6	Emergent constraints to reduce uncertainties in projections .....	75
3	5.4.7	Climate Feedbacks from CH <sub>4</sub> and N <sub>2</sub> O .....	76
4	5.4.8	Combined Biogeochemical Climate Feedback .....	77
5	5.4.9	Abrupt Changes and Tipping Points .....	78
6	5.4.9.1	Assessment of biogeochemical tipping points .....	79
7	5.4.9.1.1	Forest Dieback.....	79
8	5.4.9.1.2	Biogenic Emissions Following Permafrost Thaw .....	80
9	5.4.9.1.3	Methane Release from Clathrates .....	80
10	5.4.9.2	Abrupt Changes Detected in ESM Projections .....	81
11	5.4.10	Long Term Response past 2100 .....	81
12	5.4.11	Near-Term Prediction of Ocean and Land Carbon Sinks .....	82
13			
14	<b>5.5</b>	<b>Remaining Carbon Budgets .....</b>	<b>83</b>
15	5.5.1	Transient Climate Response to Cumulative Emissions of carbon dioxide (TCRE) .....	83
16	5.5.1.1	Contributing Physical Processes and Theoretical Frameworks.....	83
17			
18		<b>Cross-Chapter Box 5.3: The Ocean Carbon-Heat Nexus and Climate Change Commitment .....</b>	<b>84</b>
19			
20	5.5.1.2	Assessment of Limits of the TCRE Concept.....	87
21	5.5.1.2.1	Sensitivity to amount of cumulative CO <sub>2</sub> emissions.....	87
22	5.5.1.2.2	Sensitivity to the Rate of CO <sub>2</sub> Emissions .....	88
23	5.5.1.2.3	Reversibility and Earth System Feedbacks .....	88
24	5.5.1.3	Estimates of TCRE.....	89
25	5.5.1.4	Combined assessment of TCRE .....	91
26	5.5.2	Remaining Carbon Budget Assessment .....	91
27	5.5.2.1	Framework and Earlier Approaches .....	91
28	5.5.2.2	Assessment of Individual Components.....	93
29	5.5.2.2.1	TCRE .....	93
30	5.5.2.2.2	Historical Warming.....	93
31	5.5.2.2.3	Non-CO <sub>2</sub> Warming Contribution .....	94
32	5.5.2.2.4	Adjustments due to the Zero-Emission Commitment (ZEC) .....	95
33	5.5.2.2.5	Adjustments for Other not Represented Feedbacks .....	95
34	5.5.2.3	Remaining Carbon Budget.....	95
35			
36		<b>BOX 5.2: Implications of methodological advancements in estimating the remaining carbon budget</b>	
37		<b>since AR5 .....</b>	<b>97</b>
38			
39	<b>5.6</b>	<b>Biogeochemical Implications of Carbon Dioxide Removal and Solar Radiation Modification .....</b>	<b>99</b>

1	5.6.1	Introduction.....	99
2	5.6.2	Biogeochemical Responses to Carbon Dioxide Removal (CDR) .....	99
3	5.6.2.1	Global Carbon Cycle Responses to CDR.....	102
4			
5		<b>BOX 5.3: Carbon cycle response to CO<sub>2</sub> removal from the atmosphere.....</b>	<b>102</b>
6			
7	5.6.2.1.1	Carbon Cycle Response to Instantaneous CDR.....	103
8	5.6.2.1.2	Carbon Cycle Response Over Time in Scenarios with CDR .....	103
9	5.6.2.1.3	Removal Effectiveness of CDR.....	104
10	5.6.2.1.4	Symmetry of Carbon Cycle Response to Positive and Negative CO <sub>2</sub> Emissions.....	105
11	5.6.2.2	Effects of Specific CDR Methods on Biogeochemical Cycles and Climate.....	106
12	5.6.2.2.1	Land-based Biological CDR Methods .....	106
13	5.6.2.2.2	Ocean-based Biological CDR Methods .....	109
14	5.6.2.2.3	Geochemical CDR Methods.....	109
15	5.6.2.2.4	Chemical CDR methods.....	110
16	5.6.2.2.5	Methane removal .....	110
17	5.6.3	Biogeochemical responses to Solar Radiation Modification (SRM).....	111
18	5.6.3.1	Effects of SRM on the Carbon Cycle .....	111
19	5.6.3.2	Consequences of SRM and its termination on atmospheric CO <sub>2</sub> burden .....	113
20	5.6.3.3	Consequences of SRM on other Biogeochemical Cycles .....	113
21	5.6.3.4	Synthesis of biogeochemical responses to SRM.....	113
22			
23		<b>5.7 Final Remarks.....</b>	<b>113</b>
24			
25		<b>Frequently Asked Questions.....</b>	<b>116</b>
26	FAQ 5.1:	Is the natural removal of carbon from the atmosphere weakening? .....	116
27	FAQ 5.2:	Can thawing permafrost substantially increase global warming?.....	118
28	FAQ 5.3:	Could climate change be reversed by removing carbon dioxide from the atmosphere?.....	120
29	FAQ 5.4:	What are carbon budgets? .....	122
30			
31		<b>References .....</b>	<b>124</b>
32			
33		<b>Figures.....</b>	<b>177</b>
34			
35			
36			
37			
38			
39			
40			
41			

## 1 **Executive Summary**

2  
3 It is unequivocal that emissions of the well-mixed greenhouse gases (GHG) carbon dioxide (CO<sub>2</sub>), methane (CH<sub>4</sub>) and nitrous oxide (N<sub>2</sub>O) from human activities are the main driver of increases in atmospheric GHG concentrations since the pre-industrial period. The accumulation of GHGs in the atmosphere is determined by the balance between anthropogenic emissions, anthropogenic removals, and physical-biogeochemical source and sink dynamics on land and in the ocean. This chapter assesses how physical and biogeochemical processes of the carbon and nitrogen cycles affect the variability and trends of GHGs in the atmosphere as well as ocean acidification and deoxygenation. It identifies physical and biogeochemical feedbacks that have affected or could affect future rates of GHG accumulation in the atmosphere, and therefore, influence climate change and its impacts. This chapter also assesses the remaining carbon budget to limit global warming within various goals, as well as the large-scale consequences of carbon dioxide removal (CDR) and solar radiation modification (SRM) on biogeochemical cycles {Figures 5.1, 5.2}.

## 15 **The Human Perturbation of the Carbon and Biogeochemical cycles**

16  
17 **Global mean concentrations for well-mixed GHGs (CO<sub>2</sub>, CH<sub>4</sub> and N<sub>2</sub>O) in 2019 correspond to increases of about 47%, 156%, and 23%, respectively, above the levels in 1750 (representative of the pre-industrial) (*high confidence*).** Current atmospheric concentrations of the three GHGs are higher than at any point in the last 800,000 years, and in 2019 reached 409.9 ppm of CO<sub>2</sub>, 1866.3 ppb of CH<sub>4</sub>, and 332.1 ppb of N<sub>2</sub>O (*very high confidence*). Current CO<sub>2</sub> concentrations in the atmosphere are also unprecedented in the last 2 million years (*high confidence*). In the past 60 Myr, there have been periods in Earth's history when CO<sub>2</sub> concentrations were significantly higher than at present, but multiple lines of evidence show that the rate at which CO<sub>2</sub> has increased in the atmosphere during 1900–2019 is at least 10 times faster than at any other time during the last 800,000 years (*high confidence*), and 4-5 times faster than during the last 56 million years (*low confidence*). {5.1.1, 2.2.3; Figures 5.3, 5.4; Cross-Chapter Box 2.1}

## 28 **Contemporary Trends of Greenhouse Gases**

29  
30 **It is unequivocal that the increase of CO<sub>2</sub>, CH<sub>4</sub>, and N<sub>2</sub>O in the atmosphere over the industrial era is the result of human activities (*very high confidence*).** This assessment is based on multiple lines of evidence including atmospheric gradients, isotopes, and inventory data. During the last measured decade, global average annual anthropogenic emissions of CO<sub>2</sub>, CH<sub>4</sub>, and N<sub>2</sub>O, reached the highest levels in human history at  $10.9 \pm 0.9$  PgC yr<sup>-1</sup> (2010–2019), 335–383 Tg CH<sub>4</sub> yr<sup>-1</sup> (2008–2017), and 4.2–11.4 TgN yr<sup>-1</sup> (2007–2016), respectively (*high confidence*). {5.2.1, 5.2.2, 5.2.3, 5.2.4; Figures 5.6, 5.13, 5.15}.

37 **The CO<sub>2</sub> emitted from human activities during the decade of 2010–2019 (decadal average  $10.9 \pm 0.9$  PgC yr<sup>-1</sup>) was distributed between three Earth system components: 46% accumulated in the atmosphere ( $5.1 \pm 0.02$  PgC yr<sup>-1</sup>), 23% was taken up by the ocean ( $2.5 \pm 0.6$  PgC yr<sup>-1</sup>) and 31% was stored by vegetation in terrestrial ecosystems ( $3.4 \pm 0.9$  PgC yr<sup>-1</sup>) (*high confidence*).** Of the total anthropogenic CO<sub>2</sub> emissions, the combustion of fossil fuels was responsible for 81–91%, with the remainder being the net CO<sub>2</sub> flux from land-use change and land management (e.g., deforestation, degradation, regrowth after agricultural abandonment or peat drainage). {5.2.1.2, 5.2.1.5; Table 5.1; Figures 5.5, 5.7, 5.12}

46 **Over the past six decades, the average fraction of anthropogenic CO<sub>2</sub> emissions that has accumulated in the atmosphere (referred to as the airborne fraction) has remained nearly constant at approximately 44%.** The ocean and land sinks of CO<sub>2</sub> have continued to grow over the past six decades in response to increasing anthropogenic CO<sub>2</sub> emissions (*high confidence*). Interannual and decadal variability of the regional and global ocean and land sinks indicate that these sinks are sensitive to climate conditions and therefore to climate change (*high confidence*). {5.2.1.1, 5.2.1.3, 5.2.1.4.2; Figures 5.7, 5.8, 5.10}

53 **Recent observations show that ocean carbon processes are starting to change in response to the growing ocean sink, and these changes are expected to contribute significantly to future weakening of the ocean sink under medium- to high-emission scenarios.** However, the effects of these changes is not

1 yet reflected in a weakening trend of the contemporary (1960–2019) ocean sink (*high confidence*). {5.1.2,  
2 5.2.1.3, 5.3.2.1; Figures 5.8, 5.20; Cross-Chapter Box 5.3}

3  
4 **Atmospheric concentration of CH<sub>4</sub> grew at an average rate of  $7.6 \pm 2.7$  ppb yr<sup>-1</sup> for the last decade  
5 (2010–2019), with a faster growth of  $9.3 \pm 2.4$  ppb yr<sup>-1</sup> over the last six years (2014–2019) (*high*  
6 *confidence*).** The multi-decadal growth trend in atmospheric CH<sub>4</sub> is dominated by anthropogenic activities  
7 (*high confidence*), and the growth since 2007 is largely driven by emissions from both fossil fuels and  
8 agriculture (dominated by livestock) sectors (*medium confidence*). The interannual variability is dominated  
9 by El Niño–Southern Oscillation cycles, during which biomass burning and wetland emissions, as well as  
10 loss by reaction with tropospheric hydroxyl radical OH play an important role. {5.2.2; Figures 5.13, 5.14;  
11 Table 5.2; Cross-Chapter Box 5.2}

12  
13 **Atmospheric concentration of N<sub>2</sub>O grew at an average rate of  $0.85 \pm 0.03$  ppb yr<sup>-1</sup> between 1995 and  
14 2019, with a further increase to  $0.95 \pm 0.04$  ppb yr<sup>-1</sup> in the most recent decade (2010–2019).** This  
15 increase is dominated by anthropogenic emissions, which have increased by 30% between the 1980s and the  
16 most recent observational decade (2007–2016) (*high confidence*). Increased use of nitrogen fertilizer and  
17 manure contributed to about two-thirds of the increase during the 1980–2016 period, with the fossil  
18 fuels/industry, biomass burning, and wastewater accounting for much of the rest (*high confidence*). {5.2.3;  
19 Figures 5.15, 5.16, 5.17}

## 20 21 Ocean Acidification and Ocean Deoxygenation

22  
23 **Ocean acidification is strengthening as a result of the ocean continuing to take up CO<sub>2</sub> from human-  
24 caused emissions (*very high confidence*).** This CO<sub>2</sub> uptake is driving changes in seawater chemistry that  
25 result in the decrease of pH and associated reductions in the saturation state of calcium carbonate, which is a  
26 constituent of skeletons or shells of a variety of marine organisms. These trends of ocean acidification are  
27 becoming clearer globally, with a *very likely* rate of decrease in pH in the ocean surface layer of 0.016 to  
28 0.020 per decade in the subtropics and 0.002 to 0.026 per decade in subpolar and polar zones since the  
29 1980s. Ocean acidification has spread deeper in the ocean, surpassing 2000 m depth in the northern North  
30 Atlantic and in the Southern Ocean. The greater projected pH declines in CMIP6 models are primarily a  
31 consequence of higher atmospheric CO<sub>2</sub> concentrations in the Shared Socio-economic Pathways (SSPs)  
32 scenarios than their CMIP5-RCP analogues {5.3.2.2, 5.3.3.1; 5.3.4.1; Figures 5.20, 5.21}

33  
34 **Ocean deoxygenation is projected to continue to increase with ocean warming (*high confidence*).** Earth  
35 system models (ESMs) project a 32–71% greater subsurface (100–600 m) oxygen decline, depending on  
36 scenario, than reported in the Special Report on the Ocean and Cryosphere (SROCC) for the period 2080–  
37 2099. This is attributed to the effect of larger surface warming in CMIP6 models, which increases ocean  
38 stratification and reduces ventilation (*medium confidence*). There is *low confidence* in the projected reduction  
39 of oceanic N<sub>2</sub>O emissions under high emission scenarios because of greater oxygen losses simulated in  
40 ESMs in CMIP6, uncertainties in the process of oceanic N<sub>2</sub>O emissions, and a limited number of modelling  
41 studies available {5.3.3.2; 7.5}.

## 42 43 Future Projections of Carbon Feedbacks on Climate Change

44  
45 **Oceanic and terrestrial carbon sinks are projected to continue to grow with increasing atmospheric  
46 concentrations of CO<sub>2</sub>, but the fraction of emissions taken up by land and ocean is expected to decline  
47 as the CO<sub>2</sub> concentration increases (*high confidence*).** ESMs suggest approximately equal global land and  
48 ocean carbon uptake for each of the SSPs scenarios. However, the range of model projections is much larger  
49 for the land carbon sink. Despite the wide range of model responses, uncertainty in atmospheric CO<sub>2</sub> by  
50 2100 is dominated by future anthropogenic emissions rather than uncertainties related to carbon–climate  
51 feedbacks (*high confidence*). {5.4.5; Figure 5.25, 5.26}

52  
53 **Increases in atmospheric CO<sub>2</sub> lead to increases in land carbon storage through CO<sub>2</sub> fertilization of  
54 photosynthesis and increased water use efficiency (*high confidence*).** However, the overall change in land  
55 carbon also depends on land-use change and on the response of vegetation and soil to continued warming

1 and changes in the water cycle, including increased droughts in some regions that will diminish the sink  
2 capacity. Climate change alone is expected to increase land carbon accumulation in the high latitudes (not  
3 including permafrost), but also to lead to a counteracting loss of land carbon in the tropics (*medium*  
4 *confidence*, Figure 5.25). More than half of the latest CMIP6 ESMs include nutrient limitations on the  
5 carbon cycle, but these models still project increasing tropical land carbon (*medium confidence*) and  
6 increasing global land carbon (*high confidence*) through the 21st century. {5.4.1, 5.4.3, 5.4.5; Figure 5.27;  
7 Cross-Chapter Box 5.1}

8  
9 **Future trajectories of the ocean CO<sub>2</sub> sink are strongly emissions-scenario dependent (*high confidence*).**  
10 Emission scenarios SSP4-6.0 and SSP5-8.5 lead to warming of the surface ocean and large reductions of the  
11 buffering capacity, which will slow the growth of the ocean sink after 2050. Scenario SSP1-2.6 limits further  
12 reductions in buffering capacity and warming, and the ocean sink weakens in response to the declining rate  
13 of increasing atmospheric CO<sub>2</sub>. There is *low confidence* in how changes in the biological pump will  
14 influence the magnitude and direction of the ocean carbon feedback. {5.4.2, 5.4.4, Cross-Chapter Box 5.3}

15  
16 **Beyond 2100, land and ocean may transition from being a carbon sink to a source under either very**  
17 **high emissions or net negative emissions scenarios, but for different reasons.** Under very high emissions  
18 scenarios such as SSP5-8.5, ecosystem carbon losses due to warming lead the land to transition from a  
19 carbon sink to a source (*medium confidence*), while the ocean is expected to remain a sink (*high confidence*).  
20 For scenarios in which CO<sub>2</sub> concentration stabilizes, land and ocean carbon sinks gradually take up less  
21 carbon as the increase in atmospheric CO<sub>2</sub> slows down. In scenarios with moderate net negative CO<sub>2</sub>  
22 emissions and CO<sub>2</sub> concentrations declining during the 21st century (e.g., SSP1-2.6), the land sink  
23 transitions to a net source in decades to a few centuries after CO<sub>2</sub> emissions become net negative, while the  
24 ocean remains a sink (*low confidence*). Under scenarios with large net negative CO<sub>2</sub> emissions and rapidly  
25 declining CO<sub>2</sub> concentrations (e.g., SSP5-3.4-OS (overshoot)), both land and ocean switch from a sink to a  
26 transient source during the overshoot period (*medium confidence*). {5.4.10, 5.6.2.1.2; Figures 5.30, 5.33}

27  
28 **Thawing terrestrial permafrost will lead to carbon release (*high confidence*), but there is *low***  
29 ***confidence* in the timing, magnitude and the relative roles of CO<sub>2</sub> versus CH<sub>4</sub> as feedback processes.**  
30 CO<sub>2</sub> release from permafrost is projected to be 3–41 PgC per 1°C of global warming by 2100, based on an  
31 ensemble of models. However, the incomplete representation of important processes such as abrupt thaw,  
32 combined with weak observational constraints, only allow *low confidence* in both the magnitude of these  
33 estimates and in how linearly proportional this feedback is to the amount of global warming. It is *very*  
34 *unlikely* that gas clathrates in terrestrial and subsea permafrost will lead to a detectable departure from the  
35 emissions trajectory during this century. {5.4.9; Box 5.1}

36  
37 **The net response of natural CH<sub>4</sub> and N<sub>2</sub>O sources to future warming will be increased emissions**  
38 **(*medium confidence*).** Key processes include increased CH<sub>4</sub> emissions from wetlands and permafrost thaw,  
39 as well as increased soil N<sub>2</sub>O emissions in a warmer climate, while ocean N<sub>2</sub>O emissions are projected to  
40 decline at centennial time scale. The magnitude of the responses of each individual process and how linearly  
41 proportional these feedbacks are to the amount of global warming is known with *low confidence* due to  
42 incomplete representation of important processes in models combined with weak observational constraints.  
43 Models project that over the 21st century the combined feedback of 0.02–0.09 W m<sup>-2</sup> °C<sup>-1</sup> is comparable to  
44 the effect of a CO<sub>2</sub> release of 5–18 PgCeq °C<sup>-1</sup> (*low confidence*). {5.4.7, 5.4.8; Figure 5.29}

45  
46 **The response of biogeochemical cycles to the anthropogenic perturbation can be abrupt at regional**  
47 **scales, and irreversible on decadal to century time scales (*high confidence*).** The probability of crossing  
48 uncertain regional thresholds (e.g., high severity fires, forest dieback) increases with climate change (*high*  
49 *confidence*). Possible abrupt changes and tipping points in biogeochemical cycles lead to additional  
50 uncertainty in 21st century GHG concentrations, but these are *very likely* to be smaller than the uncertainty  
51 associated with future anthropogenic emissions (*high confidence*). {5.4.9}

## 52 Remaining Carbon Budgets to Climate Stabilization

53 **There is a near-linear relationship between cumulative CO<sub>2</sub> emissions and the increase in global mean**

1 **surface air temperature (GSAT) caused by CO<sub>2</sub> over the course of this century for global warming**  
2 **levels up to at least 2°C relative to pre-industrial (*high confidence*).** Halting global warming would thus  
3 require global net anthropogenic CO<sub>2</sub> emissions to become zero. The ratio between cumulative CO<sub>2</sub>  
4 emissions and the consequent GSAT increase, which is called the transient climate response to cumulative  
5 emissions of CO<sub>2</sub> (TCRE), *likely* falls in the 1.0°C–2.3°C per 1000 PgC range. The narrowing of this range  
6 compared to AR5 is due to a better integration of evidence across the science in this assessment. Beyond this  
7 century, there is *low confidence* that the TCRE remains an accurate predictor of temperature changes in  
8 scenarios of very low or net negative CO<sub>2</sub> emissions because of uncertain Earth system feedbacks that can  
9 result in further warming or a path-dependency of warming as a function of cumulative CO<sub>2</sub> emissions. {5.4,  
10 5.5.1}

11  
12 **Mitigation requirements over this century for limiting maximum warming to specific levels can be**  
13 **quantified using a carbon budget that relates cumulative CO<sub>2</sub> emissions to global mean temperature**  
14 **increase (*high confidence*).** For the period 1850–2019, a total of 655 ± 65 PgC (2390 ± 240 GtCO<sub>2</sub>) of  
15 anthropogenic CO<sub>2</sub> has been emitted. Remaining carbon budgets (starting from 1 January 2020) for limiting  
16 warming to 1.5°C, 1.7°C, and 2.0°C are 140 PgC (500 GtCO<sub>2</sub>), 230 PgC (850 GtCO<sub>2</sub>) and 370 PgC (1350  
17 GtCO<sub>2</sub>), respectively, based on the 50th percentile of TCRE. For the 67th percentile, the respective values  
18 are 110 PgC (400 GtCO<sub>2</sub>), 190 PgC (700 GtCO<sub>2</sub>) and 310 PgC (1150 GtCO<sub>2</sub>). These remaining carbon  
19 budgets may vary by an estimated ± 60 PgC (220 GtCO<sub>2</sub>) depending on how successfully future non-CO<sub>2</sub>  
20 emissions can be reduced. Since AR5 and SR1.5, estimates have undergone methodological improvements,  
21 resulting in larger, yet consistent estimates. {5.5.2, 5.6; Figure 5.31; Table 5.8}

22  
23 **Several factors affect the precise value of remaining carbon budgets, including estimates of historical**  
24 **warming, future emissions from thawing permafrost, and variations in projected non-CO<sub>2</sub> warming.**  
25 Remaining carbon budget estimates can increase or decrease by 150 PgC (550 GtCO<sub>2</sub>, *likely* range) due to  
26 uncertainties in the level of historical warming, and by an additional ± 60 PgC (±220 GtCO<sub>2</sub>, *likely* range) due  
27 to geophysical uncertainties surrounding the climate response to non-CO<sub>2</sub> emissions such as CH<sub>4</sub>, N<sub>2</sub>O, and  
28 aerosols. Permafrost thaw is included in the estimates together with other feedbacks that are often not  
29 captured by models. Despite the large uncertainties surrounding the quantification of the effects of additional  
30 Earth system feedback processes, such as emissions from wetlands and permafrost thaw, these feedbacks  
31 represent identified additional amplifying risk factors that scale with additional warming and mostly increase  
32 the challenge of limiting warming to specific temperature thresholds. These uncertainties do not change the  
33 basic conclusion that global CO<sub>2</sub> emissions would need to decline to at least net zero to halt global warming.  
34 {5.4, 5.5.2}

### 35 **Biogeochemical Implications of Carbon Dioxide Removal and Solar Radiation Modification**

36  
37  
38 **Land- and ocean-based carbon dioxide removal (CDR) methods have the potential to sequester CO<sub>2</sub>**  
39 **from the atmosphere, but the benefits of this removal would be partially offset by CO<sub>2</sub> release from**  
40 **land and ocean carbon stores (*very high confidence*).** The fraction of CO<sub>2</sub> removed that remains out of the  
41 atmosphere, a measure of CDR effectiveness, decreases slightly with increasing amount of removal (*medium*  
42 *confidence*) and decreases strongly if CDR is applied at lower CO<sub>2</sub> concentrations (*medium confidence*).  
43 {5.6.2.1; Figures 5.32, 5.33, 5.34}

44  
45 **The century-scale climate–carbon cycle response to a CO<sub>2</sub> removal from the atmosphere is not always**  
46 **equal and opposite to the response to a CO<sub>2</sub> emission (*medium confidence*).** For simultaneously  
47 cumulative CO<sub>2</sub> emissions and removals of greater than or equal to 100 PgC, CO<sub>2</sub> emissions are 4 ± 3%  
48 more effective at raising atmospheric CO<sub>2</sub> than CO<sub>2</sub> removals are at lowering atmospheric CO<sub>2</sub>. The  
49 asymmetry originates from state-dependencies and non-linearities in carbon cycle processes and implies that  
50 an extra amount of CDR is required to compensate for a positive emission of a given magnitude to attain the  
51 same change in atmospheric CO<sub>2</sub>. The net effect of this asymmetry on the global surface temperature is  
52 poorly constrained due to *low agreement* between models (*low confidence*). {5.6.2.1; Figure 5.35}

53  
54 **Wide-ranging side-effects of CDR methods have been identified that can either weaken or strengthen**  
55 **the carbon sequestration and cooling potential of these methods and affect the achievement of**



1 **sustainable development goals (*high confidence*)**. Biophysical and biogeochemical side-effects of CDR  
2 methods are associated with changes in surface albedo, the water cycle, emissions of CH<sub>4</sub> and N<sub>2</sub>O, ocean  
3 acidification and marine ecosystem productivity (*high confidence*). These side-effects and associated Earth  
4 system feedbacks can decrease carbon uptake and/or change local and regional climate, and in turn limit the  
5 CO<sub>2</sub> sequestration and cooling potential of specific CDR methods (*medium confidence*). Deployment of  
6 CDR, particularly on land, can also affect water quality and quantity, food production and biodiversity, with  
7 consequences for the achievement of related sustainable development goals (*high confidence*). These effects  
8 are often highly dependent on local context, management regime, prior land use, and scale of deployment  
9 (*high confidence*). A wide range of co-benefits are obtained with methods that seek to restore natural  
10 ecosystems or improve soil carbon (*high confidence*). The biogeochemical effects of terminating CDR are  
11 expected to be small for most CDR methods (*medium confidence*). {5.6.2.2; Figure 5.36; Cross-Chapter Box  
12 5.1}

13  
14 **Solar radiation modification (SRM) would increase the global land and ocean CO<sub>2</sub> sinks (*medium***  
15 ***confidence*) but would not stop CO<sub>2</sub> from increasing in the atmosphere, thus exacerbating ocean**  
16 **acidification under continued anthropogenic emissions (*high confidence*)**. SRM acts to cool the planet  
17 relative to unmitigated climate change, which would increase the land sink by reducing plant and soil  
18 respiration and slow the reduction of ocean carbon uptake due to warming (*medium confidence*). SRM would  
19 not counteract or stop ocean acidification (*high confidence*). The sudden and sustained termination of SRM  
20 would rapidly increase global warming, with the return of positive and negative effects on the carbon sinks  
21 (*very high confidence*) {4.6.3; 5.6.3}

22  
23

ACCEPTED VERIFIED  
SUBJECT TO FINAL EDITS

## 5.1 Introduction

The physical and biogeochemical controls of greenhouse gases (GHGs) is a central motivation for this chapter, which identifies biogeochemical feedbacks that have led or could lead to a future acceleration, slowdown or abrupt transitions in the rate of GHG accumulation in the atmosphere, and therefore of climate change. A characterisation of the trends and feedbacks lead to improved quantification for the remaining carbon budgets for climate stabilisation, and the responses of the carbon cycle to atmospheric CO<sub>2</sub> removal, which is embedded in many of the mitigation scenarios, to achieve the goals of the Paris Agreement.

Changes in the abundance of well-mixed GHGs (carbon dioxide (CO<sub>2</sub>), methane (CH<sub>4</sub>) and nitrous oxide (N<sub>2</sub>O)) in the atmosphere play a large role in determining the Earth's radiative properties and its climate in the past, the present and the future (Chapters 2, 4, 6 and 7). Since 1950, the increase in atmospheric GHGs has been the dominant cause of the human-induced climate change (Section 3.3). While the main driver of changes in atmospheric GHGs over the past 200 years relate to the direct emissions from human activities, the net accumulation of GHGs in the atmosphere is controlled by biogeochemical source-sink dynamics of carbon that exchange between multiple reservoirs on land, oceans and atmosphere. The combustion of fossil fuels and land use change for the period 1750–2019 have released an estimated  $700 \pm 75$  PgC ( $1 \text{ PgC} = 10^{15}$  g of carbon) to the atmosphere of which less than half remains in the atmosphere today (Sections 5.2.1.2; 5.2.1.5) (Friedlingstein et al., 2020). This underscores the central role of terrestrial and ocean CO<sub>2</sub> sinks in regulating its atmospheric concentration (Ballantyne et al., 2012; Li et al., 2016c; Le Quéré et al., 2018a; Ciais et al., 2019; Gruber et al., 2019a; Friedlingstein et al., 2020).

The chapter covers three dominant GHGs in the human perturbation of the Earth's radiation budget for which high quality records exist: carbon dioxide (CO<sub>2</sub>), methane (CH<sub>4</sub>) and nitrous oxide (N<sub>2</sub>O) (Figure 5.1).

Section 5.1 (this section) provides the time context on how unique current and future scenarios of GHGs atmospheric concentrations and growth rates are in the Earth's history. It also introduces the main processes involved in carbon-climate feedbacks followed by an assessment of what can be learned from the paleo record towards a better understanding of contemporary and future GHGs-climate dynamics and their response to different mitigation trajectories.

Section 5.2 covers the state of the carbon cycle and other biogeochemical cycles, and global budgets of CO<sub>2</sub>, CH<sub>4</sub> and N<sub>2</sub>O for the industrial era (since 1750). The section emphasises the last 60-year period for which high-resolution observations are available and the most recent decade for comprehensive GHGs budgets. Significant advances have taken place since the IPCC fifth assessment report (AR5), particularly in constraining the annual to decadal variability of the ocean and land carbon sources and sinks, and in revealing about the sensitivity of carbon pools to current and future climate changes. There has been an important increase in modelling capability of the three GHGs both for land and oceans, atmospheric and ocean observations, and remote sensing products that has enabled to constrain the causes of the observed trends and variability.

Section 5.3 builds on SROCC covering the change in ocean acidification due to oceanic CO<sub>2</sub> uptake across the paleo, historical periods and future projections using CMIP6, with consequences for marine life (assessed in sixth assessment report (AR6) working group II (WGII)) and biogeochemical cycles. The section also assesses changes in deoxygenation of the oceans due to warming, increased stratification of the surface ocean and slowing of the meridional overturning circulation.

Section 5.4 covers the future projections of biogeochemical cycles and their feedbacks to the climate system fully utilising the database of the concentration-driven coupled model intercomparison project phase 6 (CMIP6). Since AR5, Earth system models (ESMs) have made progress towards including more complex carbon cycle and associated biogeochemical processes that enable exploring a range of possible future carbon-climate feedbacks and their influences on the climate system. The section addresses uncertainties and limits of our models to predict future dynamics for GHG emissions trajectories, as well as new understanding on processes involved in carbon-climate feedbacks and the possibility for rapid and abrupt

1 changes brought by non-linear dynamics.

2  
3 Section 5.5 covers the development of the total and remaining carbon budgets to climate stabilisation targets  
4 and the associated transient climate response to cumulative CO<sub>2</sub> emissions. The section shows the progress  
5 made since the AR5 (IPCC, 2013a) and the 1.5°C Special Report (IPCC, 2018a), particularly on key  
6 components required to estimate the remaining carbon budget, including the transient response to cumulative  
7 emissions of CO<sub>2</sub>, the zero emission commitment, the projected non-CO<sub>2</sub> warming, and the unrepresented  
8 Earth system feedbacks.

9  
10 Section 5.6 assesses the impacts of carbon dioxide removal and solar radiation modification for the purpose  
11 of climate mitigation on the global carbon cycle building from the assessment in the IPCC Special Report on  
12 Climate Change and Land (SRCCL). It includes an overview of the major carbon dioxide removal options  
13 and potential collateral biogeochemical effects beyond those intended climate mitigation strategies. The  
14 potential capacity to deliver atmospheric reductions and the socio-economic feasibility of such options are  
15 assessed in detail in AR6 working group III (WGIII).

16  
17 Finally, Section 5.7 highlights the knowledge gaps as limits to the assessment, which would have  
18 strengthened this assessment had those gaps not existed.

19  
20  
21 **[START FIGURE 5.1 HERE]**

22  
23 **Figure 5.1: Visual abstract for Chapter 5.**

24  
25 **[END FIGURE 5.1 HERE]**

### 26 27 28 **5.1.1 The Physical and Biogeochemical Processes in Carbon-Climate Feedbacks**

29  
30 The influence of anthropogenic CO<sub>2</sub> emissions and emission scenarios on the carbon – climate system is  
31 primarily driving the ocean and terrestrial sinks as major negative feedbacks ( $\beta$ ) that determine the  
32 atmospheric CO<sub>2</sub> levels, that then drive climate feedbacks through radiative forcing ( $\gamma$ ) (Figure 5.2)  
33 (Friedlingstein et al., 2006; Jones et al., 2013b; Jones and Friedlingstein, 2020). Biogeochemical feedbacks  
34 follow as an outcome of both carbon and climate forcing on the physics and the biogeochemical processes of  
35 the ocean and terrestrial carbon cycles (Figure 5.2) (Katavouta et al., 2018; Williams et al., 2019; Jones and  
36 Friedlingstein, 2020). Together, these carbon-climate feedbacks can amplify or suppress climate change by  
37 altering the rate at which CO<sub>2</sub> builds up in the atmosphere through changes in the land and ocean sources and  
38 sinks (Figure 5.2) (Jones et al., 2013b; Raupach et al., 2014; Williams et al., 2019). These changes depend on  
39 the, often non-linear, interaction of the drivers (CO<sub>2</sub> and climate) and processes in the ocean and land as well  
40 as the emission scenarios (Figure 5.2; Sections 5.4 and 5.6) (Raupach et al., 2014; Schwinger et al., 2014;  
41 Williams et al., 2019). There is *high confidence* that carbon-climate feedbacks and their century scale  
42 evolution play a critical role in two linked climate metrics that have significant climate and policy  
43 implications: (i) the fraction of anthropogenic CO<sub>2</sub> emissions that remains in the atmosphere, the so-called  
44 airborne fraction of CO<sub>2</sub> (AF) (Figure 5.2, Section 5.2.1.2, Figure 5.7, FAQ 5.1), and (ii) the quasi-linear  
45 trend characteristic of the transient temperature response to cumulative CO<sub>2</sub> emissions (TCRE)  
46 (MacDougall, 2016; Williams et al., 2016; Jones and Friedlingstein, 2020; Section 5.5) and other GHGs  
47 (CH<sub>4</sub> and N<sub>2</sub>O). This chapter assesses the implications of these issues from a carbon cycle processes  
48 perspective (Figure 5.2) in Sections 5.2 (historical and contemporary), 5.3 (changing carbonate chemistry),  
49 5.4 (future projections), 5.5 (remaining carbon budget) and 5.6 (response to carbon dioxide removal and  
50 solar radiation modification).

51  
52 The airborne fraction is an important constraint for adjustments in carbon-climate feedbacks and reflects the  
53 partitioning of CO<sub>2</sub> emissions between reservoirs by the negative feedbacks, which for the decade 2010–  
54 2019 were 31% on land and 23% in the ocean and also dominated the historical period (Figure 5.2; Table  
55 5.1) (Friedlingstein et al., 2020). During the period 1959–2019, the airborne fraction has largely followed the

1 growth in anthropogenic CO<sub>2</sub> emissions with a mean of 44% and a large interannual variability (Ballantyne  
2 et al., 2012; Ciais et al., 2019; Friedlingstein et al., 2020) (Section 5.2.1.2; Table 5.1). The negative feedback  
3 to CO<sub>2</sub> concentrations is associated with its impact on the air-sea and air-land CO<sub>2</sub> exchange through  
4 strengthening of partial pressure of CO<sub>2</sub> (pCO<sub>2</sub>) gradients as well as the internal processes that enhance  
5 uptake. Two of these key processes are the buffering capacity of the ocean and the CO<sub>2</sub> fertilisation effect on  
6 gross primary production (Section 5.4.1–5.4.4).

7  
8 Positive and negative climate and carbon feedbacks involve, (i) fast processes on land and oceans at time  
9 scales from minutes to years such as photosynthesis, soil respiration, net primary production, shallow ocean  
10 physics and air-sea fluxes, and (ii) slower processes taking decades to millennia such as changing ocean  
11 buffering capacity, ocean ventilation, vegetation dynamics, permafrost changes, peat formation and  
12 decomposition (Figure 5.2) (Ciais et al., 2013; Forzieri et al., 2017; Williams et al., 2019). Depending on the  
13 particular combination of driver process and response dynamics, they behave as positive or negative  
14 feedbacks that amplify or dampen the magnitude and rates of climate change, respectively (Cox et al., 2000;  
15 Friedlingstein et al., 2003, 2006; Hauck and Völker, 2015; Williams et al., 2019); red and turquoise arrows  
16 in Figure 5.2; Section 5.2.1.5; Table 5.1).

17  
18 Carbon cycle feedbacks co-exist with climate (heat and moisture) feedbacks (Cross-Chapter Boxes 5.1 and  
19 5.3), which together drive contemporary (Section 5.2) and future (Section 5.4) carbon-climate feedbacks  
20 (Williams et al., 2019). The excess heat generated by radiative forcing from increasing concentration of  
21 atmospheric CO<sub>2</sub> and other GHGs is mostly taken up by the ocean (> 90%) and the residual balance  
22 partitioned between atmospheric, terrestrial and ice melting (Cross-Chapter Box 9.2; Frölicher et al., 2015).  
23 The combined effect of these two large scale negative feedbacks of CO<sub>2</sub> and heat are reflected in the TCRE  
24 (Section 5.5; Cross-Chapter Box 5.3), which points to a quasi-linear and quasi-emission-path independent  
25 relationship between cumulative emissions of CO<sub>2</sub> and global warming, which is used as the basis to  
26 estimate the remaining carbon budget (Section 5.5) (MacDougall and Friedlingstein, 2015; MacDougall,  
27 2017; Bronselaer and Zanna, 2020; Jones and Friedlingstein, 2020). There is still *low confidence* on the  
28 relative roles and importance of the ocean and terrestrial carbon processes on TCRE variability and  
29 uncertainty on centennial time scales (MacDougall, 2016; MacDougall et al., 2017; Williams et al., 2017a;  
30 Katavouta et al., 2018, 2019; Jones and Friedlingstein, 2020) (Sections 5.5.1.1, 5.5.1.2).

31  
32  
33 **[START FIGURE 5.2 HERE]**

34  
35 **Figure 5.2: Key compartments, processes and pathways that govern historical and future CO<sub>2</sub> concentrations**  
36 **and carbon–climate feedbacks through the coupled earth system.** The anthropogenic CO<sub>2</sub> emissions,  
37 including land use change, are partitioned via negative feedbacks (turquoise dotted arrows) between the  
38 ocean (23%), the land (31%) and the airborne fraction (46%) of anthropogenic CO<sub>2</sub> that sets the changing  
39 CO<sub>2</sub> concentration in the atmosphere (2010–2019, Table 5.1). This regulates most of the radiative forcing  
40 that creates the heat imbalance that drives the climate feedbacks to the ocean (blue) and land (green).  
41 Positive feedbacks (red arrows) result from processes in the ocean and on land (red text). Positive  
42 feedbacks are influenced by both carbon-concentration and carbon-climate feedbacks simultaneously.  
43 Additional biosphere processes have been included but these have an as yet uncertain feedback impact  
44 (blue-dotted arrows). CO<sub>2</sub> removal from the atmosphere into the ocean, land and geological reservoirs,  
45 necessary for negative emissions, has been included (grey arrows). Although this schematic is built  
46 around CO<sub>2</sub>, the dominant GHG, some of the same processes also influence the fluxes of CH<sub>4</sub> and N<sub>2</sub>O  
47 and the strength of the positive feedbacks from the terrestrial and ocean systems.

48  
49 **[END FIGURE 5.2 HERE]**

50  
51  
52 The combined effects of climate and CO<sub>2</sub> concentration feedbacks on the global carbon cycle are projected  
53 by ESMs to modify both the processes and natural reservoirs of carbon on a regional and global scale that  
54 may result in positive feedbacks (red arrows in Figure 5.2), which could weaken the major terrestrial and  
55 ocean sinks and disrupt both the airborne fraction and TCRE under medium to high emission scenarios  
56 (Figure 5.25; Section 5.4.5).

### 5.1.2 *Paleo Trends and Feedbacks*

Paleoclimatic proxy records extend beyond the variability of recent decadal climate oscillations and thus provide an independent perspective on feedbacks between climate and carbon cycle dynamics. These past changes according to reconstructions were slower than the current anthropogenic ones, so they cannot provide an unequivocal comparison. Nonetheless, they can help appraise sensitivities and point toward potentially dominant mechanisms of change (Tierney et al., 2020) on (sub)centennial to (multi)millennial timescales.

AR5 (WGI, Chapter 5) concluded with *medium confidence* that atmospheric CO<sub>2</sub> concentrations reached 350–450 ppm during the mid-Pliocene (3.3–3.0 Ma), and possibly 1000 ppm during the Early Eocene (52–48 Ma). AR5 (WGI, Chapter 5) also concluded with *very high confidence* that the current rates of CO<sub>2</sub>, CH<sub>4</sub> and N<sub>2</sub>O rise in atmospheric concentrations were unprecedented with respect to the ice core record covering the last deglacial transition (LDT, 18–11 ka) and with *medium confidence* that the rate of change of the reconstructed GHG rise was also unprecedented compared to the lower resolution of the records of the past 800 kyr.

#### 5.1.2.1 *Cenozoic Proxy CO<sub>2</sub> Record*

Quantifying past changes in the rate of CO<sub>2</sub> accumulation in the atmosphere based on reconstructions using marine sediment proxies is complex as age model uncertainties, assumptions and shortcomings underlying proxy applications and sedimentary processes conspire to alter and confound rate estimates (Ajayi et al., 2020). Indeed, differential sediment mixing and bioturbation contribute to smooth and attenuate proxy records (Hupp and Kelly, 2020), thereby tending to underestimate maximum rates of change (Kemp et al., 2015). Considering the extent to which uncertainties can affect sediment-based rate estimates and notwithstanding recent effort in minimizing their inherent contribution, there is generally *low to medium confidence* in quantifying rates of change on timescale less than a decade back thousands of years, and less than a millennium back millions of years in the past based on marine sediments.

In the past, atmospheric CO<sub>2</sub> concentrations reached much higher levels than present day (see Cross-Chapter Box 2.1; Figure 5.3). In particular, the Paleocene-Eocene thermal maximum (PETM), 55.9–55.7 Ma (Figure 5.3), provides some level of comparison with the current and projected anthropogenic increase in CO<sub>2</sub> emissions (Chapter 2). Atmospheric CO<sub>2</sub> concentrations increased from about 900 to around 2000 ppm in 3–20 kyr as a result of geological carbon release to the ocean-atmosphere system (Zeebe et al., 2016; Gutjahr et al., 2017; Cui and Schubert, 2018; Kirtland Turner, 2018). There is *low to medium confidence* in evaluations of the total amount of carbon released during the PETM, as proxy data constrained estimates vary from around 3000 to more than 7000 PgC, with methane hydrates, volcanic emissions, terrestrial and/or marine organic carbon, or some combination thereof, as the probable sources of carbon (Zeebe et al., 2009; Cui et al., 2011; Gutjahr et al., 2017; Luo et al., 2016; Jones et al., 2019; Elling et al., 2020; Haynes & Hönisch, 2020). Methane emissions related to hydrate/permafrost thawing and fossil carbon oxidation may have acted as positive feedbacks (Lunt et al., 2011; Armstrong McKay and Lenton, 2018; Lyons et al., 2019), as the inferred increase in atmospheric CO<sub>2</sub> can only account for approximately half of the reported warming (Zeebe et al., 2009). The estimated, time-integrated carbon input is broadly similar to the RCP8.5 extension scenario, although CO<sub>2</sub> emission rates (0.3–1.5 Pg yr<sup>-1</sup>) and by inference the rate of CO<sub>2</sub> accumulation in the atmosphere (4–42 ppm per century) during the PETM were at least 4–5 lower than during the modern era (from 1995 to 2014, Table 2.1) (Zeebe et al., 2016; Gingerich, 2019).

[START FIGURE 5.3 HERE]

**Figure 5.3: Atmospheric CO<sub>2</sub> concentrations and growth rates for the past 60 million years and projections to 2100.** (a) CO<sub>2</sub> concentrations. Concentrations data for the period 60 Myr to the time prior to 800 Kyr (left

column) are shown as the LOESS Fit and 68% range (data from Chapter 2) (Foster et al., 2017). Concentrations from 1750 and projections through 2100 are taken from Shared Socioeconomic Pathways of IPCC AR6 (Meinshausen et al., 2017). (b) Growth rates are shown as the time derivative of the concentration time series. Inserts in (b) show growth rates at the scale of the sampling resolution. Further details on data sources and processing are available in the chapter data table (Table 5.SM.6).

### [END FIGURE 5.1 HERE]

The last 50 Myr have been characterised by a gradual decline in atmospheric CO<sub>2</sub> levels at a rate of ~16 ppm Myr<sup>-1</sup> (Foster et al., 2017; Gutjahr et al., 2017) (Figure 5.3). The exact cause of this long-term change in CO<sub>2</sub> remains uncertain, but may be related to an imbalance between long-term sources of CO<sub>2</sub> (volcanic outgassing) and long-term sinks (organic carbon burial and silicate weathering).

The most recent time interval when atmospheric CO<sub>2</sub> concentration was as high as 1000 ppm (i.e. similar to the end-of 21st century projection for the high-end emission scenario RCP8.5) was around 33.5 Ma, prior to the Eocene-Oligocene transition (Zhang et al., 2013; Anagnostou et al., 2016). Atmospheric CO<sub>2</sub> levels then reached a critical threshold (1000–750 ppm, (DeConto et al., 2008)) to allow for the development of permanent regional ice-sheets on Antarctica, associated with changes in Southern Ocean hydrography, which would have increased deep ocean CO<sub>2</sub> storage (Leutert et al., 2020).

The most recent interval characterised by atmospheric CO<sub>2</sub> levels similar to modern (i.e. 360–420 ppm) was the Mid-Pliocene warm period (MPWP; 3.3–3.0 Myr, (Martínez-Botí et al., 2015a; de la Vega et al., 2020)) (Chapter 2). The relatively high atmospheric CO<sub>2</sub> concentration during the MPWP are related to vigorous ocean circulation and a rather inefficient marine biological carbon pump (Burls et al., 2017), which would have reduced deep ocean carbon storage. After the MPWP, atmospheric CO<sub>2</sub> concentrations declined gradually at a rate of 30 ppm Myr<sup>-1</sup> (de la Vega et al., 2020) (Figure 5.3), as an increase in ocean stratification led to enhanced ocean carbon storage, allowing for major, sustained advances in northern hemisphere ice sheets, 2.7 Ma (Sigman et al., 2004; DeConto et al., 2008).

#### 5.1.2.2 *Glacial-Interglacial Greenhouse Gases Records*

The Antarctic ice core record covering the past 800 kyr provides an important archive to explore the carbon-climate feedbacks prior to anthropogenic perturbations (Brovkin et al., 2016). Polar ice cores represent the only climatic archive from which past GHG concentrations can be directly measured. Major GHGs, CH<sub>4</sub>, N<sub>2</sub>O and CO<sub>2</sub> generally co-vary on orbital timescales (Loulergue et al., 2008; Lüthi et al., 2008; Schilt et al., 2010) (Chapter 2), with consistently higher atmospheric concentrations during warm intervals of the past, pointing to a strong sensitivity to climate (Figure 5.4). Modelling work suggests that the carbon cycle contributed to globalise and amplify changes in orbital forcing, which are pacing glacial-interglacial climate oscillations (Ganopolski and Brovkin, 2017), with ocean biogeochemistry and physics, terrestrial vegetation, peatland, permafrost and exchanges with the lithosphere including chemical weathering, volcanic activity, sediment burial and marine calcium carbonate compensation all playing a role in modulating the concentration of atmospheric GHGs.

Since AR5, the number of ice core records and the temporal resolution of their data for the last 800 kyr have improved, in particular for the last 60 kyr. Additionally, the advent of isotopic measurements on GHGs extracted from air trapped in ice, allows for more robust source apportionments and inventory assessments. The ensuing discussion will thus mainly focus on these two specific aspects.

Major pre-industrial sources of CH<sub>4</sub> comprise wetlands (including subglacial environments) and biomass burning (Bock et al., 2010, 2017; Lamarche-Gagnon et al., 2019; Kleinen et al., 2020). Pre-industrial atmospheric N<sub>2</sub>O concentrations were regulated by microbial production in marine and terrestrial environments and by photochemical removal in the stratosphere (Schilt et al., 2014; Battaglia and Joos, 2018; Fischer et al., 2019). Pre-industrial atmospheric CO<sub>2</sub> concentrations were largely regulated by

1 exchange with exogenic terrestrial and ocean carbon reservoirs. The imbalance between geological sources  
2 and sinks in the ocean-atmosphere-land biosphere system additionally plays an important role in modulating  
3 the air-sea partitioning of the active carbon inventory on multi-millennial timescales (Cartapanis et al.,  
4 2018).

5  
6  
7 **[START FIGURE 5.4 HERE]**

8  
9 **Figure 5.4: Atmospheric concentrations of CO<sub>2</sub>, CH<sub>4</sub> and N<sub>2</sub>O in air bubbles and clathrate crystals in ice cores (800,000 BCE to 1990 CE).** Note the variable x-axis range and tick mark intervals for the 3 columns. Ice  
10 core data is over-plotted by atmospheric observations from 1958 to present for CO<sub>2</sub>, from 1984 for CH<sub>4</sub>  
11 and from 1994 for N<sub>2</sub>O. The time-integrated, millennial-scale linear growth rates for different time  
12 periods (800,000–0 BCE, 0–1900 CE and 1900–2017 CE) are given in each panel. For the BCE period,  
13 mean rise and fall rates are calculated for the individual slopes between the peaks (interglacials) and  
14 troughs (glacial periods), which are given in the panels in left column. The data for BCE period are used  
15 from the Vostok, EPICA, Dome C and WAIS ice cores (Petit et al., 1999; Monnin, 2001; Pépin et al.,  
16 2001; Raynaud et al., 2005; Siegenthaler et al., 2005; Loulergue et al., 2008; Lüthi et al., 2008; Schilt  
17 et al., 2010a). The data after 0-yr CE are taken mainly from Law Dome ice core analysis (MacFarling  
18 Meure et al., 2006). The surface observations for all species are taken from NOAA cooperative research  
19 network (Dlugokencky and Tans, 2019), where ALT, MLO and SPO stand for Alert (Canada), Mauna  
20 Loa Observatory, and South Pole Observatory, respectively. BCE = before current era, CE = current era.  
21 Further details on data sources and processing are available in the chapter data table (Table 5.SM.6).

22  
23  
24 **[END FIGURE 5.4 HERE]**

25  
26  
27 Model-based estimates indicate that wetland CH<sub>4</sub> emissions were reduced by 24–40% during the Last  
28 Glacial Maximum (LGM) when compared to pre-industrial, while CH<sub>4</sub> emissions related to biomass burning  
29 (wildfires) decreased by 35–75% (Valdes, et al., 2005; Hoperft et al., 2017; Kleinen et al., 2020). N<sub>2</sub>O  
30 emissions decreased by about 30% during the LGM based on data-constrained model estimates (Schilt et al.,  
31 2014; Fischer et al., 2019b) owing to a combination of a weaker hydrological cycle and a generally better  
32 ventilated intermediate depth ocean relative to present, reducing (de)nitrification processes (Galbraith and  
33 Kienast, 2013; Fischer et al., 2019b).

34  
35 During past ice ages, generally colder and drier climate conditions contributed to a substantial decline of the  
36 land biosphere carbon inventory, in particular in boreal peatlands (–300 PgC) (Treat et al., 2019). Estimates  
37 assessing the glacial decrease in the global terrestrial biosphere C stock vary between –300 and –600 PgC  
38 (Ciais et al., 2012; Peterson et al., 2014; Menviel et al., 2017; Kleinen et al., 2020), possibly –850 PgC when  
39 accounting for ocean-sediment interactions and burial (Jeltsch-Thömmes et al., 2019), a considerable  
40 contraction when compared to the modern land biosphere stock. The large range of estimates reflects a yet  
41 limited understanding on how glacially perturbed nutrient fluxes and soil dynamics, as well as largely  
42 exposed shelf areas in the tropics as a result of lowered sea-level, altered carbon cycle dynamics. Recent  
43 estimates suggest deep-sea CO<sub>2</sub> storage during the last ice age exceeded modern values by as much as 750 –  
44 950 PgC (Skinner et al., 2015; Buchanan et al., 2016; Skinner et al., 2017; Anderson et al., 2019; Gottschalk  
45 et al., 2020). A combination of increased CO<sub>2</sub> solubility associated with 2–3°C lower mean oceanic  
46 temperatures (Bereiter et al., 2018), increased oceanic residence time of CO<sub>2</sub> (Skinner et al., 2017), altered  
47 oceanic alkalinity (Yu et al., 2010a; Cartapanis et al., 2018), and a generally more efficient marine biological  
48 carbon pump (BCP) (Galbraith and Jaccard, 2015; Yu et al., 2019; Galbraith and Skinner, 2020) enhanced  
49 the partition CO<sub>2</sub> into the ocean interior, although the relative contribution of each mechanism remains a  
50 matter of debate. Recent observationally constrained Earth system model results highlight that air-sea  
51 disequilibrium amplifies the effect of cooling and iron fertilisation on glacial carbon storage (Khatriwala et  
52 al., 2019).

53  
54 Ice core observations combined with model-based estimates thus reveal with *high confidence* that both  
55 terrestrial and marine CH<sub>4</sub> and N<sub>2</sub>O emissions were reduced under glacial climate conditions. Multiple lines  
56 of evidence indicate with *high confidence* that enhanced storage of remineralised CO<sub>2</sub> in the ocean interior,

1 owing to a combination of synergistic mechanisms, was sufficient to balance the removal of carbon from the  
2 atmosphere and the terrestrial biosphere reservoirs combined during the last ice age.

3  
4 Vegetation regrowth and increased precipitation in wetland regions associated with the mid-deglacial  
5 Northern Hemisphere warming (referred to as the Bølling/Allerød (B/A) warm interval, 14.7–12.7 ka), in  
6 particular in the (sub)tropics, accounts for large increases in both CH<sub>4</sub> and N<sub>2</sub>O emissions to the atmosphere  
7 (Schilt et al., 2014; Baumgartner et al., 2014; Bock et al., 2017; Fischer et al., 2019). Specifically, changes in  
8 CH<sub>4</sub> sources were steered by variations in vegetation productivity, source size area, temperatures and  
9 precipitation as modulated by insolation, local sea-level changes and monsoon intensity (Bock et al., 2017;  
10 Kleinen et al., 2020). Changes in the CH<sub>4</sub> atmospheric sink term probably only played a secondary role in  
11 modulating atmospheric CH<sub>4</sub> inventories across the LDT (Hopcroft et al., 2017; Kleinen et al., 2020)  
12 Geological emissions, related to the destabilisation of fossil (radiocarbon-dead) CH<sub>4</sub> sources buried in  
13 continental margins as a result of sudden warming appear small (Bock et al., 2017; Petrenko et al., 2017;  
14 Dyonisius et al., 2020). Stable isotope analysis on N<sub>2</sub>O extracted from Antarctic and Greenland ice reveal  
15 that marine and terrestrial emissions increased by  $0.7 \pm 0.3$  and  $1.7 \pm 0.3$  TgN, respectively, across the LDT  
16 (Fischer et al., 2019b). During abrupt Northern Hemisphere warmings, terrestrial emissions responded  
17 rapidly to the northward displacement of the Intertropical Convergence Zone (ITCZ) associated with the  
18 resumption of the Atlantic meridional overturning circulation (AMOC) (Fischer et al., 2019b). About 90% of  
19 these step increases occurred rapidly, possibly in less than 200 years (Fischer et al., 2019b). In contrast,  
20 marine emissions increased more gradually, modulated by global ocean circulation reorganisation.

21  
22 The gradual increase in atmospheric CO<sub>2</sub> across the LDT was punctuated by three centennial 10–13 ppm  
23 increments, coeval with 100–200 ppb increases in CH<sub>4</sub> (Marcott et al., 2014), reminiscent of similar  
24 oscillations reported for the last ice age associated with transient warming events (Dansgaard/Oeschger (DO)  
25 events) (Ahn and Brook, 2014; Rhodes et al., 2017; Bauska et al., 2018) as well as previous deglacial  
26 transitions (Nehrbass-Ahles et al., 2020). The rate of change in atmospheric CO<sub>2</sub> accumulation during these  
27 transient events exceed the averaged deglacial growth rates by at least 50% (Table 2.1, Figure 5.4). The early  
28 deglacial release of remineralised carbon from the ocean abyss coincided with the resumption of Southern  
29 Ocean overturning circulation (Skinner et al., 2010; Schmitt et al., 2012; Ferrari et al., 2014; Gottschalk et  
30 al., 2016, 2020; Jaccard et al., 2016; Rae et al., 2018; Moy et al., 2019) and the concomitant reduction in the  
31 global efficiency of the marine BCP, associated, in part, with dwindling iron fertilisation (Hain et al., 2010;  
32 Martinez-Garcia et al., 2014; Jaccard et al., 2016) The two subsequent pulses, centred 14.8 and 12.9 ka, are  
33 associated with enhanced air-sea gas exchange in the Southern Ocean (Li et al., 2020a), iron fertilisation in  
34 the South Atlantic and North Pacific (Lambert et al., 2021) and rapid increase in soil respiration owing to the  
35 resumption of AMOC and associated southward migration of the ITCZ (Bauska et al., 2016; Marcott et al.,  
36 2014). Indeed, rapid warming of high northern latitudes contributed to thaw permafrost, possibly liberating  
37 labile organic carbon to the atmosphere (Köhler et al., 2014; Crichton et al., 2016; Winterfeld et al., 2018;  
38 Meyer et al., 2019). Ocean surface pH reconstructions indicate that the ocean was oversaturated with respect  
39 to the atmosphere during the early, mid-LDT (Martinez-Botí et al., 2015b; Shao et al., 2019; Shuttleworth et  
40 al., 2021), suggesting that ocean sources may have been larger than terrestrial sources then. Over the course  
41 of the LDT, the decrease in northern hemisphere permafrost carbon stocks has been more than compensated  
42 by an increase in the carbon stocks of mineral soils, peatland and vegetation (Lindgren et al., 2018; Jeltsch-  
43 Thömmes et al., 2019). The land biosphere was on average a net sink for atmospheric carbon and  
44 accumulated several hundred Gt of carbon over the LDT.

45  
46 Detailed investigations reveal that Antarctic air temperatures and more generally Southern Hemisphere  
47 (30°S–60°S) proxy temperature reconstructions led the rise in *p*CO<sub>2</sub> at the onset of the LDT, 18 ka ago, by  
48 several hundred years (Shakun et al., 2012; Chowdhry Beeman et al., 2019). Atmospheric CO<sub>2</sub>, on the other  
49 hand, led reconstructed global average temperature by several centuries (Shakun et al., 2012), corroborating  
50 the importance of CO<sub>2</sub> as an amplifier of orbitally-driven warming. The phasing between Antarctic air  
51 temperature and atmospheric GHG concentration changes was nearly synchronous, yet variable, during the  
52 LDT, owing to the complex nature of the mechanisms modulating the global carbon cycle (Chowdhry  
53 Beeman et al., 2019). Mean ocean temperature reconstructions, based on noble gas extracted from Antarctic  
54 ice are closely correlated with Antarctic air temperature and *p*CO<sub>2</sub> records, emphasising the role the Southern  
55 Ocean is playing in modulating global climate variability (Bereiter et al., 2018; Baggenstos et al., 2019).



1  
2 Enhanced mid-ocean ridge magmatism and/or hydrothermal activity modulated by sea-level rise has recently  
3 been hypothesised to have contributed to the deglacial CO<sub>2</sub> rise (Crowley et al., 2015; Lund et al., 2016;  
4 Huybers and Langmuir, 2017; Stott et al., 2019b). While geological carbon release may have affected the  
5 ocean's radiocarbon budget (Ronge et al., 2016; Rafter et al., 2019; Stott et al., 2019a), model results suggest  
6 however that the potential contribution of geological carbon sources to the atmosphere remained small (Roth  
7 and Joos, 2012; Hasenclever et al., 2017).

8  
9 Simulations of Earth models of intermediate complexity (EMIC) with coupled glacial-interglacial climate  
10 and the carbon cycle were able to reproduce first-order changes in the atmospheric CO<sub>2</sub> content for the first  
11 time in recent years (Ganopolski and Brovkin, 2017; Khatiwala et al., 2019). The most important processes  
12 accounting for the full deglacial CO<sub>2</sub> amplitude in the models include solubility changes, changes in oceanic  
13 circulation and marine carbonate chemistry. The effect of the terrestrial carbon cycle, variable volcanic  
14 outgassing and the temperature dependence on the oceanic remineralisation length scale contribute less than  
15 15 ppm CO<sub>2</sub> between the glacial and interglacial intervals of the cycles. However, details in the simulated  
16 response of the marine carbon cycle and atmospheric CO<sub>2</sub> concentrations to changes in ocean circulation  
17 depend to a large degree on model parametrisation (Gottschalk et al., 2019).

18  
19 Independent paleoclimatic evidence suggests with *high confidence* that marine and terrestrial CH<sub>4</sub> and N<sub>2</sub>O  
20 emissions are highly sensitive to climate on (sub)centennial timescales. Limited, yet internally consistent ice  
21 core measurements indicate with *medium confidence* that pulsed geologic CH<sub>4</sub> release from continental  
22 margins associated with warming remained negligible across the LDT. Multiple lines of evidence suggest  
23 with *high confidence* that CO<sub>2</sub> was released from the ocean interior on centennial timescales during the LDT  
24 in response to or associated with warming, contributing to the transition out of the last glacial stage to the  
25 current interglacial period.

26  
27 Multiple lines of evidence inferred from marine sediment proxies indicate with *low to medium confidence*  
28 that the millennial rates of CO<sub>2</sub> concentration change in the atmosphere during the last 56 Myr were at least  
29 4-5 times lower than during the last century (Figure 5.3). In spite of uncertainties in ice core reconstructions  
30 related to delayed enclosure of air bubbles, which tend to smooth the records, there is *high confidence* that  
31 the rates of atmospheric CO<sub>2</sub> and CH<sub>4</sub> change during the last century were at least 10 and 5 times faster,  
32 respectively, than the maximum centennial growth rate averages of those gases during the last 800 kyr (Fig.  
33 5.4).

### 34 35 36 5.1.2.3 Holocene Changes

37  
38 Atmospheric GHG concentrations were much less variable during the pre-industrial Holocene (from 11.7 ka  
39 to 1750). Atmospheric CH<sub>4</sub> concentrations decreased at the beginning of the Holocene, consistent with a  
40 general weakening of boreal sources (Yang et al., 2017; Beck et al., 2018) and further decline during the  
41 Mid-Holocene owing to a reduction in Southern Hemisphere emissions concomitant with a southward shift  
42 of the ITCZ (Singarayer et al., 2011; Beck et al., 2018). Atmospheric CH<sub>4</sub> concentrations increased about 5  
43 ka, which prompted the hypothesis of an early anthropogenic influence, related to land use changes in  
44 southeast Asia (Ruddiman et al., 2016). However, stable isotope compositions on CH<sub>4</sub> extracted from  
45 Greenland and Antarctic ice (Beck et al., 2018) reveal that natural emissions located in the southern tropics  
46 were responsible for the rise in atmospheric CH<sub>4</sub> concentrations, in line with model simulations (Singarayer  
47 et al., 2011) thus disputing the early anthropogenic influence on the global CH<sub>4</sub> budget. Atmospheric N<sub>2</sub>O  
48 concentrations increased slightly (20 ppb) across the Holocene, associated with a gradual decline in its  
49 nitrogen stable isotope composition (Fischer et al., 2019b). The combined signal is consistent with a small  
50 increase in terrestrial emissions, offset by a reduction in marine emissions (Schilt et al., 2010b; Fischer et al.,  
51 2019b).

52  
53 The early Holocene decrease in CO<sub>2</sub> concentration by about 5 ppm (Schmitt et al., 2012) has been attributed  
54 to post-glacial regrowth in terrestrial biomass and a gradual increase in peat reservoirs over the Holocene,  
55 resulting in the sequestration of several hundred PgC (Yu et al., 2010; Nichols and Peteet, 2019). Peat

1 accumulation rates in boreal and temperate regions were higher under warmer summer conditions in the  
2 early to middle Holocene (Loisel et al., 2014; Stocker et al., 2017). The 20 ppm gradual increase of  
3 atmospheric CO<sub>2</sub> starting 7 ka has been attributed to a decrease in natural terrestrial biomass due to climate  
4 change, carbonate compensation and enhanced shallow water carbonate deposition (Menviel and Joos, 2012;  
5 Brovkin et al., 2016b), consistent with stable carbon isotope measurements on CO<sub>2</sub> extracted from Antarctic  
6 ice (Elsig et al., 2009; Schmitt et al., 2012). These isotopic measurements do not support an early  
7 anthropogenic influence on atmospheric CO<sub>2</sub> due to land use change and forest clearing (Ruddiman et al.,  
8 2016). Recent paleoceanographic evidence suggests that remineralised carbon outgassing associated with  
9 increased Southern Ocean circulation and upwelling (Studer et al., 2018), possibly promoted by stronger  
10 Southern Hemisphere westerly winds (Saunders et al., 2018), could have additionally contributed to the late  
11 Holocene increase in atmospheric CO<sub>2</sub> concentrations. However, the role of these mechanisms remained  
12 insignificant in transient Holocene ESM simulations (Brovkin et al., 2019). Overall, as in AR5 (WGI,  
13 Chapter 5), there is *medium confidence* in the key drivers of the CO<sub>2</sub> increase between the early Holocene  
14 and the beginning of the industrial era yet there is *low confidence* to the relative contributions of these  
15 drivers due to insufficient quantitative constraints on particular processes.

## 18 5.2 Historical Trends, Variability and Budgets of CO<sub>2</sub>, CH<sub>4</sub>, and N<sub>2</sub>O

19  
20 This section assesses the trends and variability in atmospheric accumulation of the three main GHGs (CO<sub>2</sub>,  
21 CH<sub>4</sub> and N<sub>2</sub>O), their ocean and terrestrial sources and sinks as well as their budgets during the Industrial Era  
22 (1750–2019). Emphasis is placed on the more recent contemporary period (1959–2019) where understanding  
23 is increasingly better constrained by atmospheric, ocean and land observations. The section also assesses our  
24 increased understanding of the anthropogenic forcing and processes driving the trends, as well as how  
25 variability at the seasonal to decadal scales provide insights on the mechanism governing long-term trends  
26 and emerging biogeochemical-climate feedbacks with their regional characteristics.

### 29 5.2.1 CO<sub>2</sub>: Trends, Variability and Budget

#### 31 5.2.1.1 Anthropogenic CO<sub>2</sub> Emissions

32  
33 There are two anthropogenic sources of CO<sub>2</sub>: fossil emissions and net emissions (including removals)  
34 resulting from land use change and land management (also shown in this chapter as LULUCF: land-use,  
35 land-use change, and forestry, termed forestry and other land use (FOLU) in previous IPCC reports). Fossil  
36 CO<sub>2</sub> emissions include the combustion of the fossil fuels coal, oil and gas covering all sectors of the  
37 economy (electricity, transport, industrial, and buildings), fossil carbonates such as in cement manufacturing,  
38 and other industrial processes such as the production of chemicals and fertilisers (Figure 5.5a). Fossil CO<sub>2</sub>  
39 emissions are estimated by combining economic activity data and emission factors, with different levels of  
40 methodological complexity (tiers) or approaches (e.g., IPCC Guidelines for National Greenhouse Gas  
41 Inventories). Several organisations or groups provide estimates of fossil CO<sub>2</sub> emissions, with each dataset  
42 having slightly different system boundaries, methods, activity data, and emission factors (Andrew, 2020).  
43 Datasets cover different time periods, which can dictate the datasets and methods that are used for a  
44 particular application. The data reported here is from an annually updated data source that combines multiple  
45 sources to maximise temporal coverage (Friedlingstein et al., 2020). The uncertainty in global fossil CO<sub>2</sub>  
46 emissions is estimated to be ±5% (1 standard deviation).

47  
48 Fossil CO<sub>2</sub> emissions have grown continuously since the beginning of the industrial era (Figure 5.5) with  
49 short intermissions due to global economic crises or social instability (Peters et al., 2012; Friedlingstein et  
50 al., 2020). In the most recent decade (2010–2019), fossil CO<sub>2</sub> emissions reached an average  $9.6 \pm 0.5$  PgC yr<sup>-1</sup>  
51 and were responsible for 86% of all anthropogenic CO<sub>2</sub> emissions during. In 2019, fossil CO<sub>2</sub> emissions  
52 were estimated to be  $9.9 \pm 0.5$  PgC yr<sup>-1</sup> excluding carbonation (Friedlingstein et al., 2020) the highest on  
53 record. These estimates excluding the cement carbonation sink of around 0.2 PgC yr<sup>-1</sup>. Fossil CO<sub>2</sub> emissions  
54 grew at 0.9% yr<sup>-1</sup> in the 1990s, increasing to 3.0% per year in the 2000s, and reduced to 1.2% from 2010 to

2019. The slower growth in fossil CO<sub>2</sub> emissions in the last decade is due to a slowdown in growth from coal use. CO<sub>2</sub> emissions from coal use grew at 4.8% yr<sup>-1</sup> in the 2000s, but they slowed to 0.4% yr<sup>-1</sup> in the 2010s. CO<sub>2</sub> emissions from oil use grew steadily at 1.1% yr<sup>-1</sup> in both the 2000s and 2010s. CO<sub>2</sub> emissions from gas use grew at 2.5% yr<sup>-1</sup> in the 2000s and 2.4% yr<sup>-1</sup> in 2010s, but is showing signs of accelerated growth of 3% yr<sup>-1</sup> since 2015 (Peters et al., 2020a). Direct CO<sub>2</sub> emissions from carbonates in cement production are around 4% of total fossil CO<sub>2</sub> emissions, and grew at 5.8% yr<sup>-1</sup> in the 2000s but a slower 2.4% yr<sup>-1</sup> in the 2010s. The uptake of CO<sub>2</sub> in cement infrastructure (carbonation) offsets about one half of the carbonate emissions from current cement production (Friedlingstein et al., 2020). These results are robust across the different fossil CO<sub>2</sub> emission datasets, despite minor differences in levels and rates as expected given the reported uncertainties (Andrew, 2020). During 2020, the COVID-19 pandemic led to a rapid, temporary decline in fossil CO<sub>2</sub> emissions, estimated to be around 7% based on a synthesis of four estimates. (Forster et al., 2020; Friedlingstein et al., 2020; Le Quéré et al., 2020; Liu et al., 2020c) (Cross-Chapter Box 6.1).

[START FIGURE 5.5 HERE]

**Figure 5.5: Global anthropogenic CO<sub>2</sub> emissions.** (a) Historical trends of anthropogenic CO<sub>2</sub> emission (fossil fuels and net land use change, including land management, called LULUCF flux in the main text) for the period 1870 to 2019, with ‘others’ representing flaring, emissions from carbonates during cement manufacture. Data sources: (Boden et al., 2017; IEA, 2017; Andrew, 2018; BP, 2018; Le Quéré et al., 2018a; Friedlingstein et al., 2020). (b) The net land use change CO<sub>2</sub> flux (PgC yr<sup>-1</sup>) as estimated by three bookkeeping models and 16 Dynamic Global Vegetation Models (DGVMs) for the global annual carbon budget 2019 (Friedlingstein et al., 2020). The three bookkeeping models are from Hansis et al., (2015); Houghton and Nassikas, (2017); Gasser et al., (2020) and are all updated to 2019; their average is used to determine the net land use change flux in the annual global carbon budget (black line). The DGVM estimates are the result of differencing a simulation with and without land use changes run under observed historical climate and CO<sub>2</sub>, following the Trendy v9 protocol; they are used to provide an uncertainty range to the bookkeeping estimates (Friedlingstein et al., 2020). All estimates are unsmoothed annual data. Estimates differ in process comprehensiveness of the models and in definition of flux components included in the net land use change flux. Further details on data sources and processing are available in the chapter data table (Table 5.SM.6).

[END FIGURE 5.5 HERE]

The global net flux from land use change and land management is composed of carbon fluxes from land use conversions, land management and changes therein (Pongratz et al., 2018) and is equivalent to the LULUCF fluxes from the agriculture, forestry and other land use (AFOLU) sector (Jia et al., 2019). It thus consists of gross emissions (loss of biomass and soil carbon in clearing or logging, harvested product decay, emissions from peat drainage and burning, degradation) and gross removals (CO<sub>2</sub> uptake in natural vegetation regrowing after harvesting or agricultural abandonment, afforestation). The LULUCF flux relates to direct human interference with terrestrial vegetation, as opposed to the natural carbon fluxes occurring due to interannual variability or trends in environmental conditions (in particular climate, CO<sub>2</sub>, and nutrient deposition) (Houghton, 2013).

Progress since AR5 and the SRCCL (IPCC, 2019a) allows more accurate estimates of gross and net fluxes due to availability of more models, model advancement in terms of inclusiveness of land-use practices (see below), and advanced land use forcings (Ciais et al., 2013; Goldewijk et al., 2017; Hurtt et al., 2020). In addition, important terminological discrepancies were resolved. First, synergistic effects of land use change and environmental changes have been identified as a key reason for the large discrepancies between model estimates of the LULUCF flux, explaining up to 50% of differences (Pongratz et al., 2014; Stocker and Joos, 2015; Gasser et al. 2020). Another reason for discrepancies relates to natural fluxes being considered as part of the LULUCF flux when occurring on managed land in the United Nation Framework Convention on Climate Change (UNFCCC) national greenhouse gas inventories; these fluxes are considered part of the natural terrestrial sink in global vegetation models and excluded in bookkeeping models (Grassi et al., 2018). LULUCF fluxes following national GHG inventories or FAO datasets, including recent estimates (Tubiello

1 et al., 2021), are thus excluded from our global assessment, but their comparison against the academic  
2 approach is available elsewhere (at the global scale (Jia et al., 2019) and European level (Petrescu et al.,  
3 2020)).

4  
5 Land-use-related component fluxes can be verified by the growing databases of global satellite-based  
6 biomass observations in combination with information on remotely-sensed land cover change, but differ  
7 from bookkeeping and modelling with Dynamic Global Vegetation Models (DGVMs) in excluding legacy  
8 emissions from pre-satellite-era land use change and land management, and neglecting soil carbon changes,  
9 and often focus on gross deforestation, not regrowth (Jia et al., 2019).

10  
11 For the decade 2010–2019, average emissions were estimated at  $1.6 \pm 0.7$  PgC yr<sup>-1</sup> (mean  $\pm$  standard  
12 deviation, 1 sigma) (Friedlingstein et al., 2020). A *likely* general upward trend since 1850 is reversed during  
13 the second part of the 20th century (Figure 5.5b). Trends since the 1980s have *low confidence* because they  
14 differ between estimates, which is related inter alia to Houghton and Nassikas (2017) using a different land  
15 use forcing than Hansis et al. (2015) and the DGVMs. Higher emissions estimates are expected from  
16 DGVMs run under transient environmental conditions compared to bookkeeping estimates, because the  
17 DGVM estimate includes the loss of additional sink capacity. Because the transient setup requires a  
18 reference simulation without land use change to separate anthropogenic fluxes from natural land fluxes,  
19 LULUCF estimates by DGVMs include the sink forests would have developed in response to environmental  
20 changes on areas that in reality have been cleared (Pongratz et al., 2014). The agricultural areas that replaced  
21 these forests have a reduced residence time of carbon, lacking woody material, and thus provide a  
22 substantially smaller additional sink over time (Gitz and Ciais, 2003). The loss of additional sink capacity is  
23 growing in particular with atmospheric CO<sub>2</sub> and increases DGVM-based LULUCF flux estimates relative to  
24 bookkeeping estimates over time (Figure 5.5).

25  
26 Gross emissions are on average 2–3 times larger than the net flux from LULUCF, increasing from an  
27 average of  $3.5 \pm 1.2$  PgC yr<sup>-1</sup> for the decade of the 1960s to an average of  $4.4 \pm 1.6$  PgC yr<sup>-1</sup> during 2010–  
28 2019 (Friedlingstein et al., 2020). Gross removals partly balance these gross emissions to yield as the sum  
29 the net flux from LULUCF and increase from  $-2.0 \pm 0.7$  PgC yr<sup>-1</sup> for the 1960s to  $-2.9 \pm 1.2$  PgC yr<sup>-1</sup>  
30 during 2010–2019. These large gross fluxes show the relevance of land management such as harvesting or  
31 rotational agriculture and the large potential to reduce emissions by halting deforestation and degradation.

32  
33 More evidence on the pre-industrial LULUCF flux has emerged since AR5 in the form of new estimates of  
34 cumulative carbon losses until today and of a better understanding of natural carbon cycle processes over the  
35 Holocene (Ciais et al., 2013). Cumulative carbon losses by land use activities since the start of agriculture  
36 and forestry (pre-industrial and industrial era) have been estimated at 116 PgC based on global compilations  
37 of carbon stocks for soils (Sanderman et al., 2017) with about 70 PgC of this occurring prior to 1750, and for  
38 vegetation as 447 PgC (inner quartiles of 42 calculations: 375–525 PgC) (Erb et al., 2018). Emissions prior  
39 to 1750 can be estimated by subtracting the post-1750 LULUCF flux from Table 5.1 from the combined soil  
40 and vegetation losses until today; they would then amount to 328 (161–501) PgC assuming error ranges are  
41 independent. A share of 353 (310–395) PgC from prior to 1800 has indirectly been suggested as the  
42 difference between net biosphere flux and terrestrial sink estimates, which is compatible with ice-core  
43 records due to a low airborne fraction of anthropogenic emissions in pre-industrial times (Erb et al., 2018)  
44 (see also Section 5.1.2.3). *Low confidence* is assigned to pre-industrial emissions estimates.

45  
46 Since AR5, evidence emerged that the LULUCF flux might have been underestimated as DGVMs include  
47 anthropogenic land cover change, but often ignore land management practices not associated with a change  
48 in land cover; land management is more widely captured by bookkeeping models through use of  
49 observation-based carbon densities (Ciais et al., 2013; Pongratz et al., 2018). Sensitivity studies show that  
50 practices such as wood and crop harvesting increase global net LULUCF emissions (Arneth et al., 2017) and  
51 explain about half of the cumulative loss in biomass (Erb et al., 2018).

#### 52 53 54 5.2.1.2 Atmosphere

1 Atmospheric CO<sub>2</sub> concentration measurements in remote locations began in 1957 at the South Pole  
 2 Observatory (SPO) and in 1958 at Mauna Loa Observatory (MLO), Hawaii, USA (Keeling, 1960) (Figure  
 3 5.6a). Since then, measurements have been extended to multiple locations around the world (Bacastow et al.,  
 4 1980; Conway et al., 1994; Nakazawa et al., 1997). In addition, high density global observations of total  
 5 column CO<sub>2</sub> measurements have been made since 2009 by dedicated greenhouse gases observing satellites  
 6 (Yoshida et al., 2013; O'Dell et al., 2018). Annual mean CO<sub>2</sub> growth rates are observed to be  $1.56 \pm 0.18$   
 7 ppm yr<sup>-1</sup> (average and range from 1 standard deviation of annual values) over the 61 years of atmospheric  
 8 measurements (1959–2019), with the rate of CO<sub>2</sub> accumulation almost tripling from an average of  $0.82 \pm$   
 9  $0.29$  ppm yr<sup>-1</sup> during the decade of 1960–1969 to  $2.39 \pm 0.37$  ppm yr<sup>-1</sup> during the decade of 2010–2019  
 10 (Chapter 2). The latter agrees well with that derived for total column (XCO<sub>2</sub>) measurements by the  
 11 greenhouse gases observing satellite (GOSAT) (Figure 5.6b). The interannual oscillations in monthly-mean  
 12 CO<sub>2</sub> growth rates (Figure 5.6b) show close relationship with the El Niño southern oscillation (ENSO) cycle  
 13 (Figure 5.6b) due to the ENSO-driven changes in terrestrial and ocean CO<sub>2</sub> sources and sinks on the Earth's  
 14 surface (Section 5.2.1.4).

15  
 16 Multiple lines of evidence unequivocally establish the dominant role of human activities in the growth of  
 17 atmospheric CO<sub>2</sub>. First, the systematic increase in the difference between the MLO and SPO records (Figure  
 18 5.6a) is caused primarily by the increase in emissions from fossil fuel combustion in industrialised regions  
 19 that are situated predominantly in the northern hemisphere (Ciais et al., 2019). Second, measurements of the  
 20 stable carbon isotope in the atmosphere ( $\delta^{13}\text{C}-\text{CO}_2$ ) are more negative over time because CO<sub>2</sub> from fossil  
 21 fuels extracted from geological storage is depleted in <sup>13</sup>C (Rubino et al., 2013; Keeling et al., 2017) (Figure  
 22 5.6c). Third, measurements of the  $\delta(\text{O}_2/\text{N}_2)$  ratio show a declining trend because for every molecule of  
 23 carbon burned, 1.17 to 1.98 molecules of oxygen (O<sub>2</sub>) is consumed (Ishidoya et al., 2012; Keeling and  
 24 Manning, 2014) (Figure 5.6d). These three lines of evidence confirm unambiguously that the atmospheric  
 25 increase of CO<sub>2</sub> is due to an oxidative process (i.e. combustion). Fourth, measurements of radiocarbon (<sup>14</sup>C–  
 26 CO<sub>2</sub>) at sites around the world (Levin et al., 2010; Graven et al., 2017; Turnbull et al., 2017) show a  
 27 continued long-term decrease in the <sup>14</sup>C/<sup>12</sup>C ratio. Fossil fuels are devoid of <sup>14</sup>C and therefore fossil-fuel-  
 28 derived CO<sub>2</sub> additions decrease the atmospheric <sup>14</sup>C/<sup>12</sup>C ratio (Suess, 1955).

29  
 30  
 31 **[START FIGURE 5.6 HERE]**

32  
 33 **Figure 5.6: Time series of CO<sub>2</sub> concentrations and related measurements in ambient air.** (a) concentration time  
 34 series and MLO-SPO difference, (b) growth rates, (c) <sup>14</sup>C and <sup>13</sup>C isotopes, and (d) O<sub>2</sub>/N<sub>2</sub> ratio. The data for  
 35 Mauna Loa Observatory (MLO) and South Pole Observatory (SPO) are taken from the Scripps  
 36 Institution of Oceanography (SIO)/University of California, San Diego (Keeling et al., 2001). The global  
 37 mean CO<sub>2</sub> are taken from NOAA cooperative network (as in Chapter 2), and GOSAT monthly-mean  
 38 XCO<sub>2</sub> time series are taken from National Institute for Environmental Studies (Yoshida et al., 2013). CO<sub>2</sub>  
 39 growth rates are calculated as the time derivative of deseasonalised time series (Nakazawa et al., 1997).  
 40 The  $\Delta(\text{O}_2/\text{N}_2)$  are expressed in per meg units (=  $(\text{FF}/\text{M}) \times 10^6$ , where FF = moles of O<sub>2</sub> consumed by  
 41 fossil-fuel burning, M =  $3.706 \times 10^{19}$ , total number of O<sub>2</sub> molecules in the atmosphere (Keeling and  
 42 Manning, 2014). The <sup>14</sup>CO<sub>2</sub> time series at Barring Head, Wellington, New Zealand (BHD) is taken from  
 43 GNS Science and NIWA (Turnbull et al., 2017). The multivariate ENSO index (MEI) is shown as the  
 44 shaded background in panel (b; warmer shade indicates El Niño). Further details on data sources and  
 45 processing are available in the chapter data table (Table 5.SM.6).

46  
 47 **[END FIGURE 5.6 HERE]**

48  
 49  
 50 Over the past six decades, the fraction of anthropogenic CO<sub>2</sub> emissions that has accumulated in the  
 51 atmosphere (referred to as airborne fraction) has remained near constant at approximately 44% (Figure 5.7)  
 52 (Ballantyne et al., 2012; Ciais et al., 2019; Gruber et al., 2019b; Friedlingstein et al., 2020). This suggests  
 53 that the land and ocean CO<sub>2</sub> sinks have continued to grow at a rate consistent with the growth rate of  
 54 anthropogenic CO<sub>2</sub> emissions, albeit with large inter-annual and sub-decadal variability dominated by the  
 55 land sinks (Figure 5.7).

1 Since AR5, an alternative observable diagnostic to the airborne fraction has been proposed to understand the  
2 trends in land and ocean sinks in response to its driving atmospheric CO<sub>2</sub> concentrations (Raupach et al.,  
3 2014; Bennedson et al., 2019). It is the sink rate which is defined as the combined ocean and land sink flux  
4 per unit of atmospheric excess of CO<sub>2</sub> above pre-industrial levels (Raupach et al., 2014). The sink rate has  
5 declined over the past six decades, which indicates that the combined ocean and land sinks are not growing  
6 as fast as the growth in atmospheric CO<sub>2</sub> (Raupach et al., 2014; Bennedson et al., 2019). Possible  
7 explanations for the sink rate decline are that the land and/or ocean CO<sub>2</sub> sinks are no longer responding  
8 linearly with CO<sub>2</sub> concentrations or that anthropogenic emissions are slower than exponential (Gloor et al.,  
9 2010; Raupach et al., 2014; Bennedson et al., 2019) (Figure 5.7; Sections 5.2.1.3, 5.2.1.4). In addition, both  
10 diagnostics are influenced by major climate modes (e.g. ENSO) and volcanic eruptions that contribute to  
11 high interannual variability (Gloor et al., 2010; Frölicher et al., 2013; Raupach et al., 2014), suggesting high  
12 sensitivity to future climate change. Uncertain land use change fluxes (Section 5.2.1.2) influence the  
13 robustness of the trends. Based on the AF, it is concluded with *medium confidence* that both ocean and land  
14 CO<sub>2</sub> sinks have grown consistent with the rising of anthropogenic emissions. Further research is needed to  
15 understand the drivers of changes in the CO<sub>2</sub> sink rate.

16  
17  
18 [START FIGURE 5.7 HERE]

19  
20 **Figure 5.7: Airborne fraction and anthropogenic (fossil fuel and land use change) CO<sub>2</sub> emissions.** Data as in  
21 Section 5.2.1.1. The multivariate ENSO index (shaded) and the major volcanic eruptions are marked  
22 along the x-axis. Further details on data sources and processing are available in the chapter data table  
23 (Table 5.SM.6).

24  
25 [END FIGURE 5.7 HERE]

### 26 27 28 5.2.1.3 Ocean Carbon Fluxes and Storage

29  
30 Since AR5 and the Special Report on Ocean and the Cryosphere (SROCC), major advances in globally  
31 coordinated ocean CO<sub>2</sub> observations (Surface Ocean CO<sub>2</sub> Atlas - SOCAT and Global Ocean Data Analysis  
32 Project - GLODAP), the harmonisation of ocean and coastal observations based products, atmospheric and  
33 oceanic inversion models and forced global ocean biogeochemical models (GOBMs) have increased the  
34 level of confidence in the assessment of trends and variability of air-sea fluxes and storage of CO<sub>2</sub> in the  
35 ocean during the historical period (1960–2018) (Ciais et al., 2013; Bakker et al., 2016; Landschützer et al.,  
36 2020, 2016; Bindoff et al., 2019; Tohjima et al., 2019; DeVries et al., 2019; Gregor et al., 2019; Gruber et  
37 al., 2019c, 2019a; Olsen et al., 2020; Friedlingstein et al., 2020; Hauck et al., 2020) (See also Supplementary  
38 Materials 5.SM.1). A major advance since SROCC is that for the first time all 6 published observational  
39 product fluxes, used in this assessment, are made more comparable using a common ocean and sea-ice cover  
40 area, integration of climatological coastal fluxes scaled to increasing atmospheric CO<sub>2</sub> and an ensemble  
41 mean of ocean fluxes calculated from three re-analysis wind products (Landschützer et al., 2014, 2020;  
42 Rödenbeck et al., 2014; Zeng et al., 2014; Denvil-Sommer et al., 2019; Gregor et al., 2019; Iida et al., 2020)  
43 Supplementary Materials 5.SM.2). From a process point of view, the ocean uptake of anthropogenic carbon  
44 is a two-step set of abiotic processes that involves the exchange of CO<sub>2</sub>, first across the air-sea boundary into  
45 the surface mixed layer, followed by its transport into the ocean interior where it is stored for decades to  
46 millennia, depending on the depth of storage (Gruber et al., 2019a). Two definitions of air-sea fluxes of CO<sub>2</sub>  
47 are used in this assessment for both observational products and models:  $S_{\text{ocean}}$  is the global mean ocean CO<sub>2</sub>  
48 sink and  $F_{\text{net}}$  denotes the net spatially varying CO<sub>2</sub> fluxes (Hauck et al., 2020). Adjustment of the mean  
49 global  $F_{\text{net}}$  for the pre-industrial sea-to-air CO<sub>2</sub> flux associated with land-to-ocean carbon flux term makes  
50  $F_{\text{net}}$  comparable to  $S_{\text{ocean}}$  (Jacobson et al., 2007; Resplandy et al., 2018a; Hauck et al., 2020).

51  
52 There are multiple lines of observational and modelling evidence that support with *high confidence* the  
53 finding that in the historical period air-sea fluxes and storage of anthropogenic CO<sub>2</sub> are largely influenced by  
54 atmospheric CO<sub>2</sub> concentrations, physical ocean processes and physico-chemical carbonate chemistry, which  
55 determines the unique properties of CO<sub>2</sub> in sea water (Wanninkhof et al., 2014; DeVries et al., 2017; Gruber

1 et al., 2019, 2019a; Hauck et al., 2020; McKinley et al., 2017; 2020; Chapter 9, Cross-Chapter Box 5.3).  
2 Here we assess three different approaches (Figures 5.8a, 5.8b and 5.9) that together provide *high confidence*  
3 that during the historical period (1960–2018) the ocean carbon sink ( $S_{\text{ocean}}$ ) and its associated ocean carbon  
4 storage have grown in response to global anthropogenic CO<sub>2</sub> emissions (Gruber et al., 2019c; Hauck et al.,  
5 2020; McKinley et al., 2020).  
6  
7

#### 8 5.2.1.3.1 Ocean Carbon Fluxes and Storage: Global Multi-Decadal Trends

9 In the first assessment approach, the mean global multidecadal (1960–2019) trends in the ocean sink ( $S_{\text{ocean}}$ )  
10 for CO<sub>2</sub> show a high degree of coherence across the 9 GOBMs and 6  $p\text{CO}_2$ -based observational product  
11 reconstructions (1987–2018), which despite a temporary slowdown (or “hiatus”) in the 1990s, is also quasi-  
12 linear over that period (Figure 5.8a) (Gregor et al., 2019; Hauck et al., 2020). This coherence between the  
13 GOBMs and observations-based reconstructions (1987–2018;  $r^2=0.85$ ) provides *high confidence* that the  
14 ocean sink ( $S_{\text{ocean}}$  in Section 5.2.1.5) evaluated from GOBMs (1960–2019) grew quasi-linearly from  $1.0 \pm$   
15  $0.3 \text{ PgC yr}^{-1}$  to  $2.5 \pm 0.6 \text{ PgC yr}^{-1}$  between the decades 1960–1969 and 2010–2019 in response to global CO<sub>2</sub>  
16 emissions (Figure 5.8a; Table 5.1; Hauck et al., 2020; Friedlingstein et al., 2020). The cumulative ocean CO<sub>2</sub>  
17 uptake ( $105 \pm 20 \text{ PgC}$ ) is 23% of total anthropogenic CO<sub>2</sub> emissions ( $450 \pm 50 \text{ PgC}$ ) for the same period  
18 (Friedlingstein et al., 2020). Notwithstanding the *high confidence* in the magnitude of the annual to decadal  
19 trends for  $S_{\text{ocean}}$  this assessment is moderated to *medium confidence* by the *low confidence* in the currently  
20 inadequately constrained uncertainties in the pre-industrial land-to-ocean carbon flux, the uncertain  
21 magnitude of winter outgassing from the Southern Ocean, and the uncertain effect of the ocean surface cool-  
22 skin, the effect of data sparsity, differences between wind products and the uncertain contribution from the  
23 changing land-ocean continuum on global and regional fluxes (Jacobson et al., 2007; Resplandy et al.,  
24 2018a; Roobaert et al., 2018; Bushinsky et al., 2019; Gloege et al., 2020; Hauck et al., 2020; Watson et al.,  
25 2020). However, both GOBMs and  $p\text{CO}_2$ -based observational products independently reveal a slowdown or  
26 “hiatus” of the ocean sink in the 1990s, which provides a valuable constraint for model verification and leads  
27 to greater confidence in the model outputs (Figure 5.8a) (Landschützer et al., 2016; Gregor et al., 2018;  
28 DeVries et al., 2019; Hauck et al., 2020). A number of studies point to the role of the Southern Ocean in the  
29 global “1990s-hiatus” in air-sea CO<sub>2</sub> fluxes but provide different process-based explanations linking ocean  
30 temperature, mixing and MOC responses to variability in large scale climate systems, wind stress and  
31 volcanic activity as well as the sensitivity of the air-sea CO<sub>2</sub> flux to small changes in the atmospheric forcing  
32 from anthropogenic CO<sub>2</sub> (Landschützer et al., 2016; DeVries et al., 2017; Bronselaer et al., 2018; Gregor et  
33 al., 2018; Gruber et al., 2019c; Keppler and Landschützer, 2019; McKinley et al., 2020; Nevison et al.,  
34 2020). Data sparsity in the Southern Ocean could also be a factor amplifying the global decadal perturbation  
35 of the 1990s (Gloege et al., 2020). Therefore, while there is *high confidence* in the 1990’s-hiatus of the  
36 global ocean sink for anthropogenic CO<sub>2</sub> and that the Southern Ocean makes an observable contribution to it,  
37 there is still *low confidence* in the attribution for the processes behind the 1990s-hiatus (5.2.1.3.2). Observed  
38 increases in the amplitude of the seasonal cycle of ocean  $p\text{CO}_2$  and reductions in the mean global buffering  
39 capacity provide *high confidence* that growing CO<sub>2</sub> sink is also beginning to drive observable large-scale  
40 changes in ocean carbonate chemistry (Jiang et al., 2019). However, there is *medium confidence* that these  
41 changes, which depending on the emissions scenario could drive future ocean feedbacks, are still too small to  
42 emerge from the historical multi-decadal observed growth rate of  $S_{\text{ocean}}$  (Figure 5.8a; Sections 5.1.2; 5.3.2,  
43 5.4.2; SROCC 5.2.2.3.2) (Bates et al., 2014; Sutton et al., 2016; Fassbender et al., 2017; Landschützer et al.,  
44 2018; Jiang et al., 2019). A recent model-based study suggests that re-emergence of previously stored  
45 anthropogenic CO<sub>2</sub> is both changing the buffering capacity of the mixed layer and reducing the ocean sink  
46 for anthropogenic CO<sub>2</sub> during the historical period (Rodgers et al., 2020). This trend is not reflected in  
47 observations-based products (Figure 5.8), so we attribute a *low confidence*.  
48

49 The second assessment approach makes use of 6 independent methods to constrain the mean decadal ocean  
50 sink over the period 1990–2019 (Figure 5.8b). This provides a multi-decadal advance on the 1990–1999  
51 decadal constraint from (Denman et al., 2007) that have been widely used as a model constraint for GOBMs  
52 used for the global carbon budget (Hauck et al., 2020). The *medium confidence* attributed of this assessment  
53 of the global multi-decadal trend (Figure 5.8a) is further supported by the broad agreement in magnitude and  
54 trend of the decadal mean ocean CO<sub>2</sub> uptake with assessments that also include additional observations-  
55 based, independent methods such as ocean CO<sub>2</sub> inversion and atmospheric CO<sub>2</sub> and O<sub>2</sub>/N<sub>2</sub> measurements

(Figure 5.8b; Supplementary Materials Tables 5.SM.1 and 5.SM.2).

[START FIGURE 5.8 HERE]

**Figure 5.8: Multi-decadal trends for the ocean sink of CO<sub>2</sub>.** (a): The multi-decadal (1960–2019) trends in the annual ocean sink ( $S_{\text{ocean}}$ ) reconstructed from 9 Global Ocean Biogeochemical Models (GOBM) forced with observed atmospheric re-analysis products (Hauck et al., 2020), 6 observationally-based gap-filling products that reconstructed spatial and temporal variability in the ocean CO<sub>2</sub> flux from sparse observations of surface ocean  $p\text{CO}_2$  (Supplementary Materials 5.SM.2). The trends in  $S_{\text{ocean}}$  were calculated from the mean annual global GOBM outputs and the observational products were used to provide confidence in the GOBM assessments ( $r^2=0.85$ ). Thick lines represent the multi-model mean. Observationally-based products have been corrected for pre-industrial river carbon fluxes ( $0.62 \text{ PgC yr}^{-1}$ ) based on the average of estimates from Jacobson et al., (2007) and Resplandy et al. (2018). (b): Mean decadal constraints and their confidence intervals for global ocean sink ( $S_{\text{ocean}}$ ) of anthropogenic CO<sub>2</sub> using multiple independent or quasi-independent lines of evidence or methods for the period 1990–2019 (See Supplementary Materials Tables 5.SM.1 and 5.SM.2 for magnitudes, uncertainties and published sources). Further details on data sources and processing are available in the chapter data table (Table 5.SM.6).

[END FIGURE 5.8 HERE]

Here we provide a third comparative assessment approach depicting the spatial coherence of ocean air-sea fluxes and storage rates of CO<sub>2</sub> as well as a quantitative assessment of both fluxes for the same period (1994–2007) (Figure 5.9). Observation-based  $p\text{CO}_2$  flux products show that emissions of natural CO<sub>2</sub> occur mostly in the tropics and high latitude Southern Ocean, and that the uptake and storage of anthropogenic CO<sub>2</sub> occur predominantly in the mid-latitudes (Figure 5.9; Chapter 9; Cross-Chapter Box 5.3). Strong ocean CO<sub>2</sub> sink regions are those in the mid-latitudes associated with the cooling of poleward flowing sub-tropical surface waters as well as equatorward flowing sub-polar surface waters both of which contribute to the formation of Mode, Intermediate and Deep water masses that transport anthropogenic CO<sub>2</sub> into the ocean interior on time scales of decades to centuries in both hemispheres (Figure 5.9) (DeVries, 2014; Gruber et al., 2019a; Wu et al., 2020; Chapter 9.2.2.3). The mean decadal scale magnitude and uncertainties of  $S_{\text{ocean}}$  from net air sea fluxes ( $F_{\text{net}}$ ) were calculated from an ensemble of 6 observational based product reconstructions (Figure 5.9a) and the storage rates in the ocean interior derived from multiple ocean interior CO<sub>2</sub> data sets (Gruber et al., 2019a) (Figure 5.9b). The cumulative CO<sub>2</sub> stored in the ocean interior from 1800 to 2007 has been estimated at  $140 \pm 18 \text{ PgC}$  (Gruber et al., 2019a). As reported in SROCC (Section 5.2.2.3.1; IPCC, 2019b), the net ocean CO<sub>2</sub> storage between 1994–2007 was  $29 \pm 4 \text{ PgC}$ , which corresponds to a mean storage of  $26 \pm 5\%$  of anthropogenic CO<sub>2</sub> emissions for that period (Gruber et al., 2019a). The resulting net annual storage rate of anthropogenic CO<sub>2</sub>, equivalent to  $S_{\text{ocean}}$  for the period mid-1994–mid-2007 is  $2.2 \pm 0.3 \text{ PgCyr}^{-1}$ , which is in very close agreement with the top-down air-sea flux estimate of  $S_{\text{ocean}}$  of  $2.1 \pm 0.5 \text{ PgC yr}^{-1}$  from GOBMs and  $1.9 \pm 0.3 \text{ PgC yr}^{-1}$  from  $p\text{CO}_2$ -based observational products with the steady river carbon flux correction of  $0.62 \text{ PgCyr}^{-1}$  for the same time period (Gruber et al., 2019b; Hauck et al., 2020). This close agreement between these independent ocean CO<sub>2</sub> sink estimates derived from air-sea fluxes and storage rates in the ocean interior support the *medium confidence* assessment that the ocean anthropogenic carbon storage rates continue to be determined by the ocean sink ( $S_{\text{ocean}}$ ) in response to growing CO<sub>2</sub> emissions (McKinley et al., 2020) (Figure 5.9).

[START FIGURE 5.9 HERE]

**Figure 5.9: Comparative regional characteristics of the mean decadal (1994–2007) sea-air CO<sub>2</sub> flux ( $F_{\text{net}}$ ) and ocean storage of anthropogenic CO<sub>2</sub>.** (a) regional sink–source characteristics for contemporary ocean air-sea CO<sub>2</sub> fluxes ( $F_{\text{net}}$ ) derived from the ensemble of 6 observation-based products using SOCATv6 observational data set (Bakker et al., 2020; Landschützer et al., 2014; Rödenbeck et al., 2014; Zeng et al., 2014; Denvil-Sommer et al., 2019; Gregor et al., 2019; Iida et al., 2020). Warm colours depict outgassing fluxes and black contours characterise the the five basin-scale biomes aggregated from the original 17



1 from Fay and McKinley (2014) and adjusted by Gregor et al., (2019) also used to calculate the regional  
2 variability in flux anomalies (Supplementary Materials Figure 5.SM.1); and (b) the regional  
3 characteristics of the storage fluxes of CO<sub>2</sub> in the ocean interior for the same period (Gruber et al.,  
4 2019a). The dots reflect ocean areas where the 1-sigma standard deviation of F<sub>net</sub> from the 6 observation  
5 product is larger than the magnitude of the mean. This reflects source-sink transition areas where the  
6 mean F<sub>net</sub> is small and more strongly influenced by spatial and temporal variability across the products.  
7 Further details on data sources and processing are available in the chapter data table (Table 5.SM.6).  
8

9 **[END FIGURE 5.9 HERE]**

#### 10 5.2.1.3.2 Ocean Carbon Fluxes and Storage: Regional – Global Variability

11 The intent of this assessment is to show how global variability can be regionally forced (Gregor et al., 2019;  
12 Landschützer et al., 2019; Hauck et al., 2020). Since AR5 and SROCC, advances in global ocean CO<sub>2</sub> flux  
13 products, GOBMs and atmospheric inversion models have strengthened the confidence in the assessment of  
14 how ocean regions influence mean global variability and trends of ocean CO<sub>2</sub> air-sea fluxes (F<sub>net</sub>; see  
15 Supplementary Materials Figure 5.SM.1) (Ciais et al., 2013; Landschützer et al., 2014, 2015; Rödenbeck et  
16 al., 2014; McKinley et al., 2017; Bindoff et al., 2019; Gregor et al., 2019; Friedlingstein et al., 2020; Hauck  
17 et al., 2020). The coherence in the regional variability of the anomalies in F<sub>net</sub> from 3 independent lines of  
18 evidence support with *high confidence* that the non-steady state global interannual-decadal variability of F<sub>net</sub>  
19 has clear regional influences (Gregor et al., 2019; Landschützer et al., 2019). The tropical oceans contribute  
20 the most to the global mean interannual variability (Supplementary Materials Figure 5.SM.1d). The high  
21 latitude oceans, particularly the Southern Ocean, contribute the most to the global-scale decadal variability  
22 (Supplementary Materials Figure 5.SM.1b, c; (Landschützer et al., 2016, 2019; Gregor et al., 2019; Gruber  
23 et al., 2019c; Hauck et al., 2020). The influence of the Southern Ocean on the global mean decadal  
24 variability and the 1990's hiatus is supported by the highest regional-global correlation coefficients  
25 (Supplementary Materials Figures 5.SM.1a, c). In contrast, the Equatorial oceans' influence on global mean  
26 F<sub>net</sub> has a low correlation because, notwithstanding the coherence in interannual variability, it does not show  
27 the same global mean trend of strengthening sink in response to growing global emissions (Supplementary  
28 Materials Figure 5.SM.1d; (Gregor et al., 2019). All regions, except the Equatorial ocean, contribute to  
29 varying extents to the multidecadal trend of growth in the global ocean sink (Supplementary Materials  
30 Figure 5.SM.1). Data sparseness in the high latitudes and the relatively short length of the observational  
31 records leads to *low confidence* in the attribution of the processes that link regional-global variability to  
32 climate (Landschützer et al., 2019; Gloege et al., 2020).  
33  
34  
35

36 Regional decadal-scale anomalies in the variability of ocean CO<sub>2</sub> storage have also emerged, probably  
37 associated with changes in the meridional overturning circulation (MOC), which may influence the global  
38 variability in F<sub>net</sub> (DeVries et al., 2017; Chapter 9). In the interior of the Indian and Pacific sectors of the  
39 Southern Ocean, and the North Atlantic, the increase in the CO<sub>2</sub> inventory from 1994 to 2007 was about 20%  
40 smaller than expected from the atmospheric CO<sub>2</sub> increase during the same period and the anthropogenic CO<sub>2</sub>  
41 inventory in 1994 (Sabine, 2004; Gruber et al., 2019a). There is *medium confidence* that the ocean CO<sub>2</sub>  
42 inventory strengthened again in the decade 2005–2015 (DeVries et al., 2017). In the North Atlantic, a low  
43 rate of anthropogenic CO<sub>2</sub> storage at  $1.9 \pm 0.4$  PgC per decade during the time period of 1989–2003  
44 increased to  $4.4 \pm 0.9$  PgC per decade during 2003–2014, associated with changing ventilation patterns  
45 driven by the North Atlantic Oscillation (Woosley et al., 2016). In the Pacific sector of the Southern Ocean,  
46 the rate of anthropogenic CO<sub>2</sub> storage also increased from  $8.8 \pm 1.1$  ( $1\sigma$ ) PgC per decade during 1995–2005  
47 to  $11.7 \pm 1.1$  PgC per decade during 2005–2015 (Carter et al., 2019). However, in the Subantarctic Mode  
48 Water of the Atlantic sector of the Southern Ocean, the storage rate of the anthropogenic CO<sub>2</sub> was rather  
49 lower after 2005 than before (Tanhua et al., 2017; Bindoff et al., 2019; Section 9.2.3.2). These changes have  
50 been predominantly ascribed to the impact of changes in the MOC on the transport of anthropogenic CO<sub>2</sub>  
51 into the ocean interior due to regional climate variability, in addition to the increase in the atmospheric CO<sub>2</sub>  
52 concentration (Wanninkhof et al., 2010; Pérez et al., 2013; DeVries et al., 2017, 2019; Gruber et al., 2019b;  
53 McKinley et al., 2020) (Section 9.2.3.1). However, the low frequency of carbon observations in the interior  
54 of the vast ocean leads to *medium confidence* in the assessment of temporal variability in the rate of regional  
55 ocean CO<sub>2</sub> storage and its controlling mechanisms.  
56

1 In summary, multiple lines of both observational and modelling evidence provide *high confidence* that the  
2 finding that the ocean sink for anthropogenic CO<sub>2</sub> has increased quasi-linearly over the past 60 years in  
3 response to growing global emissions of anthropogenic CO<sub>2</sub> with a mean fraction of 23% of total emissions.  
4 The *high confidence* assessment is moderated to *medium confidence* because of a number of as yet to be  
5 adequately unconstrained ocean CO<sub>2</sub> flux terms. Observed changes in the variability of ocean pCO<sub>2</sub> and  
6 observed reductions in the mean global buffering capacity provide *high confidence* that the growing CO<sub>2</sub>  
7 sink is also beginning to drive observable large-scale changes in ocean carbonate chemistry. However, there  
8 is *medium confidence* that these changes, which depending on the emissions scenario could drive future  
9 ocean feedbacks, are still too small to emerge from the historical multi-decadal observed growth rate of  
10  $S_{\text{ocean}}$ .

#### 13 5.2.1.4 Land CO<sub>2</sub> Fluxes: Historical and Contemporary Variability and Trends

##### 15 5.2.1.4.1 Trend in Land-Atmosphere CO<sub>2</sub> Exchange

16 The global net land CO<sub>2</sub> sink is assessed to have grown over the past six decades (Ciais et al., 2019; Le  
17 Quére et al., 2018a; Sarmiento et al., 2010; Friedlingstein et al., 2019) (*high confidence*). Estimated as the  
18 residual from the mass balance budget of fossil fuel CO<sub>2</sub> emissions minus atmospheric CO<sub>2</sub> growth and the  
19 ocean CO<sub>2</sub> sink, the global net land CO<sub>2</sub> sink (including both land CO<sub>2</sub> sink and net land use change  
20 emission) increased from  $0.3 \pm 0.6 \text{ PgC yr}^{-1}$  during the 1960s to  $1.8 \pm 0.8 \text{ PgC yr}^{-1}$  during 2010s  
21 (Friedlingstein et al., 2020). An increasing global net land CO<sub>2</sub> sink since 1980s (Figure 5.10) was  
22 consistently suggested both by atmospheric inversions (e.g. Peylin et al., 2013) and by Dynamic Global  
23 Vegetation Models (e.g. Sitch et al., 2015; Friedlingstein et al., 2019). The northern hemisphere contributes  
24 more to the net increase in the land CO<sub>2</sub> sink compared to the southern hemisphere (Ciais et al., 2019), and  
25 boreal and temperate forests probably contribute the most (Tagesson et al., 2020). Attributing an increased  
26 net land CO<sub>2</sub> sink to finer regional scales remains challenging, but inversions of satellite-based column CO<sub>2</sub>  
27 products that have emerged since AR5 are a promising tool to further constrain regional land-atmosphere  
28 CO<sub>2</sub> exchange (Ciais et al., 2013; Houweling et al., 2015; Reuter et al., 2017; O'Dell et al., 2018; Palmer et  
29 al., 2019a).

31 Carbon uptake by vegetation photosynthesis exerts a first-order control over the net land CO<sub>2</sub> sink. Several  
32 lines of evidence show enhanced vegetation photosynthesis over the past decades (*medium to high*  
33 *confidence*) (Figure 5.10), including increasing satellite-derived vegetation greenness (e.g. Mao et al., 2016;  
34 Zhu et al., 2016; Jia et al., 2019; See also Chapter 2) and satellite-derived photosynthesis indicators (e.g.  
35 Badgley et al., 2017; Zhang et al., 2018b), change in atmospheric concentration of carbonyl sulphide  
36 (Campbell et al., 2017), enhanced seasonal CO<sub>2</sub> amplitude (Graven et al., 2013; Forkel et al., 2016).  
37 observation-driven inference of increasing photosynthesis CO<sub>2</sub> uptake based mostly on enhanced water use  
38 efficiency (Cheng et al., 2017), and DGVM simulated increase of photosynthesis CO<sub>2</sub> uptake (Anav et al.,  
39 2015).

41 Substantial progress has been made since AR5 on attributing change of the global net land CO<sub>2</sub> sink.  
42 Increasing global net land CO<sub>2</sub> sink since the 1980s is mainly driven by the fertilisation effect from rising  
43 atmospheric CO<sub>2</sub> concentrations (Schimel et al., 2015; Sitch et al., 2015; Fernández-Martínez et al., 2019;  
44 O'Sullivan et al., 2019; Tagesson et al., 2020; Walker et al., 2020) (*medium confidence*). Increasing nitrogen  
45 deposition (de Vries et al., 2009; Devaraju et al., 2016; Huntzinger et al., 2017) or the synergy between  
46 increasing nitrogen deposition and atmospheric CO<sub>2</sub> concentration (O'Sullivan et al., 2019) could have also  
47 contributed to the increasing global net land CO<sub>2</sub> sink. The effects of climate change alone on the global net  
48 land CO<sub>2</sub> sink is so divergent that even the signs of the effects are not the same across DGVMs (e.g.  
49 Huntzinger et al., 2017).

51 Lower fire emission of CO<sub>2</sub> and enhanced vegetation carbon uptake due to reduced global burned area have  
52 contributed to the increasing global net land CO<sub>2</sub> sink in the recent decade (Arora and Melton, 2018; Yin et  
53 al., 2020) (*low to medium confidence*). Satellite observations reveal a declining trend in global burned area  
54 by about 20% over past two decades (Andela et al., 2017; Earl and Simmonds, 2018; Forkel et al., 2019b;  
55 Jones et al., 2020), a trend most pronounced in regions like northern Africa (Forkel et al., 2019; Zubkova et

1 al., 2019; Bowman et al., 2020; Jones et al., 2020) and Mediterranean Europe (Turco et al., 2016). However,  
 2 burned area trends are highly heterogeneous regionally with increasing trends reported in regions like  
 3 western United States (Holden et al., 2018a; Abatzoglou et al., 2019) and southeastern Australia (Canadell et  
 4 al., 2020). Some regions (e.g. Amazon basin and Australia) experienced record-breaking fire events in 2019  
 5 and 2020 (e.g. Boer et al., 2020), whose effects on burned area trends remain to be explored. The burned  
 6 area trends were primarily attributed to both human-induced climate change and human activities (Jolly et  
 7 al., 2015; Andela et al., 2017; Holden et al., 2018b; Turco et al., 2018; Teckentrup et al., 2019; Bowman et  
 8 al., 2020), as well as changing frequency of lightning in the boreal region (Veraverbeke et al., 2017). In  
 9 addition to changes in the burned area, fire dynamics could affect trend in land-atmosphere CO<sub>2</sub> exchange  
 10 indirectly through increasing concentration of air pollutants (Yue and Unger, 2018; Lasslop et al., 2019; see  
 11 Section 6.3.4 for impacts of ozone and aerosol on the carbon cycle).

12  
 13 Significant uncertainties remain on land CO<sub>2</sub> sink partition of processes due to challenges in reconciling  
 14 multiple-scale evidence from experiments to the globe (Fatichi et al., 2019; Walker et al., 2020), due to large  
 15 spatial and inter-model differences in diagnosing dominant driving factors affecting the net land CO<sub>2</sub> sink  
 16 (Huntzinger et al., 2017; Fernández-Martínez et al., 2019), and due to model deficiency in process  
 17 representations (He et al., 2016). Nitrogen dynamics, a major gap in DGVMs identified in AR5, have now  
 18 been incorporated in about half of the DGVMs contributing to the carbon budget of the Global Carbon  
 19 Project (GCP) (see Le Quéré et al., (2018a) for model characteristics) and a growing number of ESMS  
 20 (Arora et al., 2020). However, as the representations of carbon-nitrogen interactions vary greatly among  
 21 models, large uncertainties remain on how nitrogen cycling regulates the response of ecosystem carbon  
 22 uptake to higher atmospheric CO<sub>2</sub> (Davies-Barnard et al., 2020; Walker et al., 2015; Wieder et al., 2019;  
 23 Meyerholt et al., 2020; see Section 5.4.1). Fire modules have been incorporated into 10 of 16 DGVMs  
 24 contributing to the global carbon budget (Le Quéré et al., 2018a), and a growing number of models have  
 25 representations of human ignitions and fire suppression processes (Rabin et al., 2017; Teckentrup et al.,  
 26 2019). There are also growing DGVM developments to include management practices (Pongratz et al.,  
 27 2018b) and the effects of secondary forest regrowth (Pugh et al., 2019a), though models still under represent  
 28 intensively managed ecosystems, such as croplands and managed forests (Guanter et al., 2014; Thurner et al.,  
 29 2017). Processes that have not yet played a significant role in the land CO<sub>2</sub> sink of the past decades but can  
 30 grow in importance, include permafrost (Section 5.4) and peatlands dynamics (Dargie et al., 2017; Gibson et  
 31 al., 2019), have also been incorporated in some DGVMs (Koven et al., 2015a; Burke et al., 2017a;  
 32 Guimberteau et al., 2018). Growing numbers and varieties of Earth observations are being jointly used to  
 33 drive and benchmark models, helping further identify key processes missing or mechanisms poorly  
 34 represented in the current generation of DGVMs (e.g. Collier et al., 2018).

35  
 36  
 37 **[START FIGURE 5.10 HERE]**

38  
 39 **Figure 5.10: Trends of the net land CO<sub>2</sub> sink and related vegetation observations during 1980–2019.** (a) Net land  
 40 CO<sub>2</sub> sink. The residual net land CO<sub>2</sub> sink is estimated from the global CO<sub>2</sub> mass balance (fossil fuel  
 41 emission minus atmospheric CO<sub>2</sub> growth rate and ocean CO<sub>2</sub> sink). Inversions indicate the net land CO<sub>2</sub>  
 42 sink estimated by an ensemble of four atmospheric inversions. Dynamic Global Vegetation Models  
 43 (DGVMs) indicate the mean net land CO<sub>2</sub> sink estimated by 17 dynamic global vegetation models driven  
 44 by climate change, rising atmospheric CO<sub>2</sub>, land use change and nitrogen deposition change (for carbon-  
 45 nitrogen models). The positive values indicate net CO<sub>2</sub> uptake from the atmosphere. (b) Normalised  
 46 difference vegetation index (NDVI). The anomaly of global area-weighted NDVI observed by AVHRR  
 47 and MODIS satellite sensors. AVHRR data are accessible during 1982–2016 and MODIS data are  
 48 accessible during 2000–2018. (c) Near-infrared reflectance of vegetation (NIRv) and contiguous solar-  
 49 induced chlorophyll fluorescence (CSIF). The standardised anomaly of area-weighted NIRv during  
 50 2001-2018 (Badgley et al., 2017) and CSIF during 2000-2018 (Zhang et al., 2018). (d) Gross primary  
 51 production (GPP). The GPP from Cheng et al. (2017), DGVMs and MODIS GPP product (MOD17A3).  
 52 GPP from Cheng et al. (2017) is based on an analytical model driven by climate change, rising  
 53 atmospheric CO<sub>2</sub>, AVHRR leaf area index datasets and evapotranspiration datasets. GPP from DGVMs is  
 54 the ensemble mean global GPP estimated by the same 17 DGVMs that provide the net land CO<sub>2</sub> sink  
 55 estimates. Shaded area indicates 1- $\sigma$  inter-model spread except for atmospheric inversions, whose ranges  
 56 were used due to limited number of models. Further details on data sources and processing are available  
 57 in the chapter data table (Table 5.SM.6).

1  
2 [END FIGURE 5.10 HERE]  
3  
4

#### 5.2.1.4.2 Interannual variability in land-atmosphere CO<sub>2</sub> exchange

6 AR5 stated that the interannual variability of the atmospheric CO<sub>2</sub> growth rate is dominated by tropical land  
7 ecosystems. A set of new satellite measurements applied to assess the variability of the tropical land carbon  
8 balance since AR5 (Ciais et al., 2013) confirm this statement, including satellite column CO<sub>2</sub> measurements,  
9 estimating recent anomalous land-atmosphere CO<sub>2</sub> exchange induced by El Niño at continental scale (e.g.  
10 Liu et al., 2017; Palmer et al., 2019), and L-band vegetation optical depth, estimating tropical above-ground  
11 biomass carbon stock changes (Fan et al., 2019). In addition, based on medium evidence and medium  
12 agreement between studies with DGVMs and atmospheric inversions, semi-arid ecosystems over the tropical  
13 zones have a larger contribution to interannual variability in global land-atmosphere CO<sub>2</sub> exchange than  
14 moist tropical forest ecosystems (Poulter et al., 2014; Ahlstrom et al., 2015; Piao et al., 2020) (*low to*  
15 *medium confidence*).  
16

17 Understanding the mechanisms driving interannual variability in the carbon cycle has the potential to provide  
18 insights into whether and to what extent the carbon cycle can affect the climate (carbon-climate feedback),  
19 with particular interests over the highly climate-sensitive tropical carbon cycle (e.g. Cox et al., 2013a; Fang  
20 et al., 2017; Humphrey et al., 2018; Jung et al., 2017a; Malhi et al., 2018; Wang et al., 2014; see Section  
21 5.4). Consistent findings from studies with atmospheric inversions, satellite observations and DGVMs (e.g.  
22 Malhi et al., 2018; Rödenbeck et al., 2018) lead to *high confidence* that the tropical net land CO<sub>2</sub> sink is  
23 reduced under warmer and drier conditions, particularly during El Niño events. Interannual variations in  
24 tropical land-atmosphere CO<sub>2</sub> exchange are significantly correlated with anomalies of tropical temperature,  
25 water availability and terrestrial water storage (Wang et al., 2014; Jung et al., 2017; Humphrey et al., 2018;  
26 Piao et al., 2020), whose relative contribution are difficult to separate due to covariations between these  
27 climatic factors. At continental scale, the dominant climatic driver of interannual variations of tropical land-  
28 atmosphere CO<sub>2</sub> exchange was temperature variations (Figure 5.11; Piao et al., 2020), which could partly  
29 result from the spatial compensation of the water availability effects on land-atmospheric CO<sub>2</sub> exchange  
30 (Jung et al., 2017).  
31  
32

33 [START FIGURE 5.11 HERE]  
34

35 **Figure 5.11: Interannual variation in detrended anomalies of the net land CO<sub>2</sub> sink and land surface air**  
36 **temperature during 1980–2019.** Correlation coefficients between the net land CO<sub>2</sub> sink anomalies and  
37 temperature anomalies are shown on the right bar plots. The net land CO<sub>2</sub> sink is estimated by four  
38 atmospheric inversions (blue) and fifteen Dynamic Global Vegetation Models (DGVMs) (green),  
39 respectively (Friedlingstein et al., 2020). Solid blue and green lines show model mean detrended  
40 anomalies of the net land CO<sub>2</sub> sink. The ensemble mean of DGVMs is bounded by the 1- $\sigma$  inter-model  
41 spread in each large latitude band (North 30°N–90°N, Tropics 30°S–30°N, South 90°S–30°S) and the  
42 globe. The ensemble mean of atmospheric inversions is bounded by model spread. For each latitudinal  
43 band, the anomalies of the net land CO<sub>2</sub> sink and temperature (orange) were obtained by removing the  
44 long-term trend and seasonal cycle. A 12-month running mean was applied to reduce high-frequency  
45 noise. The bars in the right panels show correlation coefficients between the net land CO<sub>2</sub> sink anomalies  
46 and temperature anomalies for each region. Two asterisks indicate P<0.01, and one indicates P<0.05.  
47 Grey shaded area shows the intensity of El Niño southern oscillation (ENSO) as defined by the Niño 3.4  
48 index. Two volcanic eruptions (El Chichón eruption and Pinatubo eruption) are indicated with light blue  
49 dashed lines. Temperature data are from the Climatic Research Unit (CRU), University of East Anglia  
50 (Harris et al., 2014). Anomalies were calculated following Patra et al. (2005), but using 12-month low-  
51 pass filter and detrended to obtain interannual variations. Further details on data sources and processing  
52 are available in the chapter data table (Table 5.SM.6).  
53

54 [END FIGURE 5.11 HERE]  
55

56 [START CROSS-CHAPTER BOX 5.1]  
57

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### **Cross-Chapter Box 5.1: Interactions between the carbon and water cycles, particularly under drought conditions**

**Contributors:** Josep G Canadell (Australia), Philippe Ciais (France), Hervé Douville (France), Sabine Fuss (Germany), Robert B Jackson (USA), Annalea Lohila (Finland), Shilong Piao (China), Sonia I. Seneviratne (Switzerland), Sergio M. Vicente-Serrano (Spain), Sönke Zaehle (Germany)

This box presents an assessment of interactions between the carbon and water cycles that influence the dynamics of the biosphere and its interaction with the climate system. It also highlights carbon-water trade-offs arising from the use of land-based climate mitigation options. Individual aspects of the interactions between the carbon and water cycles are addressed in separate chapters (Sections 5.2.1, 5.4.1, 8.2.3, 8.3.1, 8.4.1, 11.6). The influence of wetlands and dams on methane emissions is assessed elsewhere (Sections 5.2.2, 5.4.7, 8.3.1), as well as the consequences of permafrost thawing (Box 5.1, Section 9.5.2) and/or increased flooding (Sections 8.4.1, 11.5, 12.4) on wetland extent in the northern high latitudes and wet tropics.

#### **Does elevated CO<sub>2</sub> alleviate the impacts of drought?**

Increasing atmospheric CO<sub>2</sub> concentration enhances leaf photosynthesis and drives a partial closure of leaf stomata, leading to higher water-use efficiency (WUE) at the leaf, canopy to ecosystem scales (Norby and Zak, 2011; De Kauwe et al., 2013; Fatichi et al., 2016; Knauer et al., 2017; Mastrotheodoros et al., 2017). Since AR5 (Box 6.3), a growing body of evidence from tree-ring and carbon isotopes further confirms an increase of plant water-use efficiency over decadal to centennial time scales, with some evidence for a stronger enhancement of photosynthesis compared to stomatal reductions (Frank et al., 2015; Guerrieri et al., 2019; Adams et al., 2020).

Multiple lines of evidence suggest that WUE has increased in near proportionality to atmospheric CO<sub>2</sub> (*high confidence*), at a rate generally consistent with Earth System Models (ESMs), despite variation in the WUE response to CO<sub>2</sub> (De Kauwe et al., 2013; Frank et al., 2015; Keeling et al., 2017; Lavergne et al., 2019; Walker et al., 2020). Both field-scale CO<sub>2</sub> enrichment experiments and process models show the effect of physiologically induced water savings, particularly under water limiting conditions (De Kauwe et al., 2013; Farrior et al., 2015; Lu et al., 2016; Roy et al., 2016). Plants can also benefit from reduced drought stress due to enhanced CO<sub>2</sub> without ecosystems-scale water savings (Jiang et al., 2021). This increased WUE offsets to some extent the effects of enhanced vapor pressure deficit (VPD) on plant transpiration (Bobich et al., 2010; Creese et al., 2014; Jiao et al., 2019), but will have limited effect on ameliorating plant water stress during extreme drought events (Xu et al., 2016; Menezes-Silva et al., 2019; Liu et al., 2020b), when leaf stomata is governed primarily by soil moisture (Roy et al., 2016).

Leaf stomata closure can have large effects on land freshwater availability because of reduced plant transpiration leading in some regions to higher soil moisture and runoff (Roderick et al., 2015; Milly and Dunne, 2016; Yang et al., 2019c). However, increased water availability is often not realized because other CO<sub>2</sub> physiological effects that enhance ecosystem evapotranspiration might offset the gains. These effects include plant growth and leaf area expansion (Ainsworth and Long, 2005; Ukkola et al., 2016; McDermid et al., 2021), lengthening of the vegetative growing season (Frank et al., 2015; Lian et al., 2021), and the effects of stomatal closure on near-surface atmosphere that leads to increased air temperature and vapor-pressure deficits (Berg et al., 2016; Vogel et al., 2018; Zhou et al., 2019; Grossiord et al., 2020).

ESMs show no consensus about the net hydrological response to physiological CO<sub>2</sub> effects. Some studies show water savings as a consequence of the CO<sub>2</sub> effects on leaf stomata closure (Swann et al., 2016; Lemordant et al., 2018), while other studies show that increased leaf area offsets the gains from increased WUE (Mankin et al., 2019). However, these projections are subject to ESM uncertainties to quantify transpiration (Lian et al., 2021), among them the correct representations of plant hydraulic architecture such as changes in xylem anatomical properties and deep rooting (Nie et al., 2013; Liu et al., 2020b).

1 In conclusion, it is *very likely* that elevated CO<sub>2</sub> leads to increased WUE at the leaf level concurrent with  
2 enhanced photosynthesis. Increased CO<sub>2</sub> concentrations alleviate the effects of water deficits on plant  
3 productivity (*medium confidence*) but there is *low confidence* on its role under extreme drought conditions.  
4 There is *low confidence* that increased WUE by vegetation will substantially reduce global plant  
5 transpiration and diminish the frequency and severity of soil moisture and streamflow deficits associated  
6 with the radiative effect of higher CO<sub>2</sub> concentrations.

### 7 8 **How does drought affect the terrestrial CO<sub>2</sub> sink?**

9  
10 Water availability controls the spatial distribution of photosynthesis (gross primary productivity, GPP) over a  
11 larger part of the globe (Beer et al., 2010) and, at local scale, drought decreases GPP more than respiration  
12 (Schwalm et al., 2012) over most ecosystem types. This makes water availability a major climatic driver of  
13 variability in net ecosystem exchange (Jung et al., 2017; Humphrey et al., 2018). In addition to suppressing  
14 photosynthesis, field evidence suggests that droughts reduce the land CO<sub>2</sub> sink also through increasing forest  
15 mortality and promoting wildfire (Allen et al., 2015; Brando et al., 2019; Abram et al., 2021).

16  
17 At the global scale, interannual variability in the atmospheric CO<sub>2</sub> growth rate and global-scale terrestrial  
18 water storage from satellite shows that a lower global net land CO<sub>2</sub> sink is associated with below-average  
19 terrestrial water storage (Humphrey et al., 2018). Atmospheric inversions based on surface and satellite  
20 column CO<sub>2</sub> measurements show significant carbon release during drought events in pan-tropic areas  
21 (Phillips et al., 2009; Gatti et al., 2014; Liu et al., 2017a; Palmer et al., 2019a). Regional extreme droughts in  
22 the mid-latitudes also lead to decreased GPP and land CO<sub>2</sub> sink (Ciais et al., 2005; Wolf et al., 2016; Peters  
23 et al., 2020b; Flach et al., 2021). Droughts cannot be compensated by equivalent wet anomalies because of  
24 the non-linear response of the terrestrial carbon uptake to soil moisture (Green et al., 2019).

25  
26 Uncertainties remain on the magnitude of sensitivity of the land carbon fluxes to droughts. Global studies  
27 indicate stronger control of soil moisture to variations in satellite proxies of GPP than VPD (Stocker et al.,  
28 2019; Liu et al., 2020b). However, given that VPD increases exponentially with atmospheric warming, some  
29 studies suggest that the importance of VPD in stomatal regulation will become increasingly more important  
30 under warmer climate (Novick et al., 2016; Grossiord et al., 2020). It is difficult to isolate the relative  
31 contributions of warmer temperature, higher VPD and lower soil moisture. This is because land-atmosphere  
32 feedbacks cause a simultaneous increase of plant evaporative demand and of root zone water deficit  
33 impairing plant root uptake (Berg et al., 2016). These physiological responses can be further compounded by  
34 drought legacies (Anderegg et al., 2015), changes in structure and population dynamics due to forest  
35 mortality (McDowell et al., 2020), disturbances associated with drought (fire, insects damage) (Anderegg et  
36 al., 2020) and possible trade-offs between resistance and resilience (Li et al., 2020b). Nonetheless, ESMs  
37 suggest that increased drought effects under very high levels of global warming (ca. 4°C at the end of the  
38 21st century) contribute to the reduced efficiency of the land sink (Green et al., 2019).

39  
40 In conclusion, there is *high confidence* that the global net land CO<sub>2</sub> sink is reduced on interannual scale when  
41 regional-scale reductions in water availability associated with droughts occur, particularly in tropical regions.  
42 There is also *high confidence* that the global land sink will become less efficient due to soil moisture  
43 limitations and associated drought conditions in some regions for high emission scenarios, specially under  
44 global warming above 4°C. However, there is *low confidence* on how these water cycle feedbacks will play  
45 out in lower emission scenarios (at 2°C global warming or lower) due to uncertainties in regional rainfall  
46 changes and the balance between the CO<sub>2</sub> fertilisation effect, through WUE, and the radiative impacts of  
47 greenhouse gases.

### 48 49 **What are the limits of carbon dioxide removal from a water cycle perspective?**

50 Carbon dioxide removal (CDR) options based on terrestrial carbon sinks will require the appropriation of  
51 significant amounts of water at the landscape level. Most mitigation pathways that seek to limit global  
52 warming to 1.5°C or less than 2°C require the removal of about 30 to 300 GtC from the atmosphere by 2100  
53 (Rogelj et al., 2018b). Bioenergy with carbon capture and storage (BECCS), and afforestation/reforestation  
54 are the dominant CDR options used in climate stabilisation scenarios implying large requirements for land  
55 and water (Section 5.6; (Beringer et al., 2011; Boysen et al., 2017b; Fajardy and Mac Dowell, 2017; Jans et

al., 2018; Séférian et al., 2018b; Yamagata et al., 2018; Stenzel et al., 2019). A review of freshwater requirements for irrigating biomass plantations shows a range between 15 and 1250 km<sup>3</sup> per GtC<sup>-1</sup> of biomass harvest. This is equivalent to a water requirement of 99–8250 km<sup>3</sup> for the median BECCS deployment of around 3.3 GtCyr<sup>-1</sup> (Smith et al., 2016) in <2°C-scenarios (Stenzel et al., 2021), assuming that biomass is converted to electricity which is substantially less efficient than to heat. These large ranges are the result of different assumptions about the type of biomass and yield improvements, management, and land availability. The use of alternative feedstocks, such as wastes, residues and algae, would lead to smaller water requirements (Smith et al., 2019).

Most of the water consumed in BECCS is used to grow the feedstock, with carbon capture and storage constituting a smaller portion across all crops (Rosa et al., 2020), with an estimated evaporative loss of 260 km<sup>3</sup> yr<sup>-1</sup> for 3.3 GtC yr<sup>-1</sup> (Smith et al., 2016). The same authors also estimate water use for CDR through afforestation at 1040 km<sup>3</sup> yr<sup>-1</sup> for 3.3 GtC yr<sup>-1</sup>, including interception and transpiration and adjusted for the original land cover's water use.

The impacts of different CDR options on the water cycle depend crucially on regional climate, prior land cover, and scale of deployment (Trabucco et al., 2008). Extensive irrigation for afforestation in drier areas will have larger downstream impacts than in wetter regions with the difference in water use between the afforested landscapes and its previous vegetation determining the level of potential impacts on evapotranspiration and runoff (Jackson et al., 2005; Teuling et al., 2017). Afforestation and reforestation sometimes enhances precipitation through atmospheric feedbacks such as increased convection, at least in the tropics (Ellison et al., 2017) and the increase in precipitation can in some regions even cancel out the increased evapotranspiration (Li et al., 2018).

In conclusion, extensive deployment of BECCS and afforestation/reforestation will require larger amounts of freshwater resources than used by the previous vegetation, altering the water cycle at regional scales (*high confidence*). Consequences of high water consumption on downstream uses, biodiversity, and regional climate depend on prior land cover, background climate conditions, and scale of deployment (*high confidence*). Therefore, a regional approach is required to determine the efficacy and sustainability of CDR projects.

**[END CROSS-CHAPTER BOX 5.1]**

#### 5.2.1.5 CO<sub>2</sub> Budget

The global CO<sub>2</sub> budget (Figure 5.12) encompasses all natural and anthropogenic CO<sub>2</sub> sources and sinks. Table 5.1 shows the perturbation of the global carbon mass balance between reservoirs since the beginning of the industrial era, circa 1750.

Since AR5 (Ciais et al., 2013), a number of improvements have led to a more constrained carbon budget presented here. Some new additions include: (i) the use of independent estimates for the residual carbon sink on natural terrestrial ecosystems (Le Quéré et al., 2018a), (ii) improvements in the estimates of emissions from cement production (Andrew, 2019) and the sink associated with cement carbonation (Cao et al., 2020), (iii) improved and new emission estimates from forestry and other land use (Hansis et al., 2015; Gasser et al., 2020), (iv) the use of ocean observation-based sink estimates and a revised river flux partition between hemisphere (Friedlingstein et al., 2020); and (v) the expansion of constraints from atmospheric inversions, both based on surface networks and the use of satellite retrievals.

The budget, based on the annual assessment by the GCP (Friedlingstein et al., 2020), uses independent estimates of all major flux components: fossil fuel and carbonate emissions ( $E_{\text{FOS}}$ ), CO<sub>2</sub> fluxes from Forestry and other Land Use ( $E_{\text{LULUCF}}$ ), the growth rate of CO<sub>2</sub> in the atmosphere ( $G_{\text{atm}}$ ), and the ocean ( $S_{\text{ocean}}$ ) and natural land ( $S_{\text{land}}$ ) CO<sub>2</sub> sinks. An imbalance term ( $B_{\text{imb}}$ ) is required to ensure mass balance of the source and sinks that have been independently estimated:  $E_{\text{FOS}} + E_{\text{LULUCF}} = G_{\text{atm}} + S_{\text{ocean}} + S_{\text{land}} + B_{\text{imb}}$ . All estimates are reported with 1 standard deviation ( $\pm 1\sigma$ , 1 sigma) representing a likelihood of 68%.

1 Over the past decade (2010–2019),  $10.9 \pm 0.9$  PgC yr<sup>-1</sup> were emitted from human activities which were  
 2 distributed between three Earth system components: 46% accumulated in the atmosphere ( $5.1 \pm 0.02$  PgC yr<sup>-1</sup>),  
 3 23% was taken up by the ocean ( $2.5 \pm 0.6$  PgC yr<sup>-1</sup>) and 31% was stored by vegetation in terrestrial  
 4 ecosystems ( $3.4 \pm 0.9$  PgC yr<sup>-1</sup>) (Table 5.1). There is a budget imbalance of 0.1 PgCyr<sup>-1</sup> which is within the  
 5 uncertainties of the other terms. Over the industrial era (1750–2019), the total cumulative CO<sub>2</sub> fossil fuel and  
 6 industry emissions were  $445 \pm 20$  PgC, and the LULUCF flux (=net land use change in Figure 5.12) was  $240$   
 7  $\pm 70$  PgC (*medium confidence*). The equivalent total emissions ( $685 \pm 75$  PgC) was distributed between the  
 8 atmosphere ( $285 \pm 5$  PgC), oceans ( $170 \pm 20$  PgC) and land ( $230 \pm 60$  PgC) (Table 5.1), with a budget  
 9 imbalance of 20 PgC.

10  
 11 This budget (Table 5.1) does not explicitly account for source/sink dynamics due to carbon cycling in the  
 12 land–ocean aquatic continuum comprising freshwaters, estuaries, and coastal areas. Natural and  
 13 anthropogenic transfers of carbon from soils to freshwater systems are significant ( $2.4$ – $5.1$  PgC yr<sup>-1</sup>)  
 14 (Regnier et al., 2013a; Drake et al., 2018). Some of the carbon is buried in freshwater bodies ( $0.15$  PgC)  
 15 (Mendonça et al., 2017), and a significant proportion returns to the atmosphere via outgassing from lakes,  
 16 rivers and estuaries (Raymond et al., 2013; Regnier et al., 2013a; Lauerwald et al., 2015). The net export of  
 17 carbon from the terrestrial domain to the open oceans is estimated to be  $0.80$  PgC yr<sup>-1</sup> (*medium confidence*),  
 18 based on the average of (Jacobson et al., 2007; Resplandy et al., 2018b) and corrected to account for  $0.2$  PgC  
 19 buried in ocean floor sediments. These terms are included in Figure 5.12. Inclusion of other smaller fluxes  
 20 could further constrain the carbon budget (Ito, 2019; Friedlingstein et al., 2020).

21  
 22  
 23 [START FIGURE 5.12 HERE]

24  
 25 **Figure 5.12: Global carbon (CO<sub>2</sub>) budget (2010–2019).** Yellow arrows represent annual carbon fluxes (in PgC yr<sup>-1</sup>)  
 26 associated with the natural carbon cycle estimated for the time prior to the industrial era, around 1750.  
 27 Pink arrows represent anthropogenic fluxes averaged over the period 2010–2019. The rate of carbon  
 28 accumulation in the atmosphere is equal to net land-use change emissions, including land management  
 29 (called LULUCF in the main text) plus fossil fuel emissions, minus land and ocean net sinks (plus a small  
 30 budget imbalance, Table 5.1). Circles with yellow numbers represent pre-industrial carbon stocks in PgC.  
 31 Circles with pink numbers represent anthropogenic changes to these stocks (cumulative anthropogenic  
 32 fluxes) since 1750. Anthropogenic net fluxes are reproduced from Friedlingstein et al., (2020). The  
 33 relative change of *gross photosynthesis* since pre-industrial times is based on 15 DGVMs used in  
 34 Friedlingstein et al., (2020). The corresponding emissions by *Total respiration and fire* are those required  
 35 to match the *net land flux*, exclusive of net land-use change emissions which are accounted for separately.  
 36 The cumulative change of anthropogenic carbon in the terrestrial reservoir is the sum of carbon  
 37 cumulatively lost by net land use change emissions, and net carbon accumulated since 1750 in response  
 38 to environmental drivers (warming, rising CO<sub>2</sub>, nitrogen deposition). The adjusted gross natural ocean-  
 39 atmosphere CO<sub>2</sub> flux was derived by rescaling the value in figure 1 of (Sarmiento and Gruber, 2002) of  
 40  $70$  PgC/yr by the revised estimate of the bomb <sup>14</sup>C inventory in the ocean. The original bomb <sup>14</sup>C  
 41 inventory yielded an average global gas transfer velocity of  $22$  cm/hr; the revised estimate is  $17$ cm/h  
 42 leading to  $17/22*70=54$ . Dissolved organic carbon reservoir and fluxes from (Hansell et al., 2009).  
 43 Dissolved inorganic carbon exchanges between surface and deep ocean, subduction and obduction from  
 44 (Levy et al., 2013) Levy et al. 2013. Export production and flux from (Boyd et al., 2019). NPP and  
 45 remineralisation in surface layer of the ocean from (Kwiatkowski et al., 2020; Séférian et al., 2020). Deep  
 46 ocean reservoir from (Keppeler et al., 2020). Note that the mass balance of the two ocean carbon stocks  
 47 *surface ocean* and *intermediate and deep ocean* includes a yearly accumulation of anthropogenic carbon  
 48 (not shown). Fossil fuel reserves are from (BGR, 2019); fossil fuel resources are  $11,490$  PgC for coal,  
 49  $6,780$  PgC for oil, and  $365$  PgC for natural gas. Permafrost region stores are from (Hugelius et al., 2014;  
 50 Strauss et al., 2017; Mishra et al., 2021) (see also Box 5.1) and soil carbon stocks outside of permafrost  
 51 region from Batjes, (2016); Jackson et al., (2017). Biomass stocks (range of seven estimates) are from  
 52 Erb et al., (2018). Sources for the fluxes of the continuum land-to-ocean are provided in main text and  
 53 adjusted within the ranges of the various assessment to balance the budget (section 5.2.1.5).

54  
 55 [END FIGURE 5.12 HERE]

56  
 57  
 58 [START TABLE 5.1 HERE]

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**Table 5.1: Global anthropogenic CO<sub>2</sub> budget accumulated since the industrial revolution (onset in 1750) and averaged over the 1980s, 1990s, 2000s, and 2010s.** By convention, a negative ocean or land to atmosphere CO<sub>2</sub> flux is equivalent to a gain of carbon by these reservoirs. The table does not include natural exchanges (e.g. rivers, weathering) between reservoirs. Uncertainties represent the 68% confidence interval (Friedlingstein et al., 2020).

	1750–2019 Cumulative (PgC)	1850–2019 Cumulative (PgC)	1980–1989 Mean Annual Growth Rate (PgC yr <sup>-1</sup> )	1990–1999 Mean Annual Growth Rate (PgC yr <sup>-1</sup> )	2000–2009 Mean Annual Growth Rate (PgC yr <sup>-1</sup> )	2010–2019 Mean Annual Growth Rate (PgC yr <sup>-1</sup> )
<b>Emissions</b>						
Fossil fuel combustion and cement production	445 ± 20	445 ± 20	5.4 ± 0.3	6.3 ± 0.3	7.7 ± 0.4	9.4 ± 0.5
Net land use change	240 ± 70	210 ± 60	1.3 ± 0.7	1.4 ± 0.7	1.4 ± 0.7	1.6 ± 0.7
Total emissions	685 ± 75	655 ± 65	6.7 ± 0.8	7.7 ± 0.8	9.1 ± 0.8	10.9 ± 0.9
<b>Partition</b>						
Atmospheric increase	285 ± 5	265 ± 5	3.4 ± 0.02	3.2 ± 0.02	4.1 ± 0.02	5.1 ± 0.02
Ocean sink <sup>c</sup>	170 ± 20	160 ± 20	1.7 ± 0.4	2.0 ± 0.5	2.1 ± 0.5	2.5 ± 0.6
Terrestrial sink	230 ± 60	210 ± 55	2.0 ± 0.7	2.6 ± 0.7	2.9 ± 0.8	3.4 ± 0.9
<b>Budget imbalance</b>	0	20	-0.4	-0.1	0	-0.1

[END TABLE 5.1 HERE]

### 5.2.2 CH<sub>4</sub>: Trends, Variability and Budget

Methane is a much more powerful greenhouse gas than CO<sub>2</sub> (Chapter 7) and participates in tropospheric air chemistry (Chapter 6). The CH<sub>4</sub> variability in the atmosphere is mainly the result of the net balance between the sources and sinks on the Earth's surface and chemical losses in the atmosphere. Atmospheric transport evens out the regional CH<sub>4</sub> differences between different parts of the Earth's atmosphere. The steady-state lifetime is estimated to be 9.1 ± 0.9 (Chapter 6, Section 3.1). About 90% of the loss of atmospheric CH<sub>4</sub> occurs in the troposphere by reaction with hydroxyl (OH) radical, 5% by bacterial soil oxidation, and the rest 5% by chemical reactions with OH, excited state oxygen (O<sup>1</sup>D), and atomic chlorine (Cl) in the stratosphere (Saunio et al., 2020). Methane has large emissions from both natural and anthropogenic origins, but a clear demarcation of their nature is difficult because of the use and conversions of the natural ecosystem for human activities. The largest natural sources are from wetlands, freshwater and geological process, while the largest anthropogenic emissions are from enteric fermentation and manure treatment, landfills and waste treatment, rice cultivation and fossil fuel exploitation (Table 5.2). In the past two centuries, CH<sub>4</sub> emissions have nearly doubled, predominantly human driven since 1900, and persistently exceeded the losses (*virtually certain*), thereby increasing the atmospheric abundance as evidenced from the ice core and firn air measurements (Ferretti et al., 2005; Ghosh et al., 2015).

This section discusses both bottom-up and top-down estimates of emissions and sinks. Bottom-up estimates are based on empirical upscaling of point measurements, emission inventories and dynamical model simulations, while top-down estimates refer to those constrained by atmospheric measurements and chemistry-transport models in inversion systems. Since the AR5, a larger suite of atmospheric inversions using both in situ and remote sensing measurement have led to better understanding of the regional CH<sub>4</sub> sources (Cross-Chapter Box 5.2). New ice core measurements of <sup>14</sup>C-CH<sub>4</sub> are used for estimating the geological sources of CH<sub>4</sub> (Table 5.2). Compared to the IPCC SRCCL (IPCC, 2019a; Jia et al., 2019), we provide a whole atmospheric sources-sinks budget consisting of all emissions and losses.

#### 5.2.2.1 Atmosphere

Since the start of direct measurements of CH<sub>4</sub> in the atmosphere in the 1970s (Figure 5.13), the highest

1 growth rate was observed from 1977 to 1986 at  $18 \pm 4$  ppb yr<sup>-1</sup> (multi-year mean and 1 standard deviation)  
 2 (Rice et al., 2016). This rapid CH<sub>4</sub> growth followed the green revolution with increased crop-production and  
 3 a fast rate of industrialisation that caused rapid increases in CH<sub>4</sub> emissions from ruminant animals, rice  
 4 cultivation, landfills, oil and gas industry and coal mining (Ferretti et al., 2005; Ghosh et al., 2015; Crippa et  
 5 al., 2020). Due to increases in oil prices in the early 1980s, emissions from gas flaring declined significantly  
 6 (Stern and Kaufmann, 1996). This explains the first reduction in CH<sub>4</sub> growth rates from 1985 to 1990 (Steele  
 7 et al., 1992; Chandra et al., 2021). Further reductions in emission occurred following the Mt Pinatubo  
 8 eruption in 1991 that triggered a reduction in CH<sub>4</sub> growth rate through a decrease in wetland emissions  
 9 driven by lower surface temperatures due to the scattering aerosols (Bândă et al., 2016; Chandra et al., 2021).  
 10 In the late 1990s through 2006 there was a temporary pause in the CH<sub>4</sub> growth rate, with higher confidence  
 11 on its causes than the in AR5: emission from the oil and gas sectors declined by about 10 Tg yr<sup>-1</sup> through the  
 12 1990s, and atmospheric CH<sub>4</sub> loss steadily increased (Dlugokencky et al., 2003; Simpson et al., 2012; Crippa  
 13 et al., 2020; Höglund-Isaksson et al., 2020; Chandra et al., 2021). Methane growth rate began to increase  
 14 again at  $7 \pm 3$  ppb yr<sup>-1</sup> during 2007–2016, the causes of which are highly debated since the AR5 (Rigby et  
 15 al., 2008; Dlugokencky et al., 2011; Dalsøren et al., 2016; Nisbet et al., 2016; Patra et al., 2016; Schaefer et  
 16 al., 2016; Schwietzke et al., 2016; Turner et al., 2017; Worden et al., 2017; He et al., 2020); studies disagree  
 17 on the relative contribution of thermogenic, pyrogenic and biogenic emission processes and variability in  
 18 tropospheric OH concentration. The renewed CH<sub>4</sub> increase is accompanied by a reversal of  $\delta^{13}\text{C}$  trend to  
 19 more negative values post 2007; opposite to what occurred in the 200 years prior (Ferretti et al., 2005; Ghosh  
 20 et al., 2015; Schaefer et al., 2016; Schwietzke et al., 2016; Nisbet et al., 2019), suggesting an increasing  
 21 contribution from animal farming, landfills and waste, and a slower increase in emissions from fossil fuel  
 22 exploitation since the early 2000s (Patra et al., 2016; Jackson et al., 2020; Chandra et al., 2021). A  
 23 comprehensive assessment of the CH<sub>4</sub> growth rates over the past 4 decades is presented in the Cross-Chapter  
 24 Box 5.2.

25  
 26  
 27 **[START FIGURE 5.13 HERE]**

28  
 29 **Figure 5.13: Time series of CH<sub>4</sub> concentrations, growth rates and isotopic composition.** a) CH<sub>4</sub> concentrations, b)  
 30 CH<sub>4</sub> growth rates, c)  $\delta^{13}\text{C}$ -CH<sub>4</sub>. Data from selected site networks operated by NOAA (Dlugokencky et al.,  
 31 2003), AGAGE (Prinn et al., 2018) and PDX (Portland State University) (Rice et al., 2016). To maintain  
 32 clarity, data from many other measurement networks are not included here, and all measurements are  
 33 shown in WMO X2004ACH<sub>4</sub> global calibration standard. Global mean values of XCH<sub>4</sub> (total-column),  
 34 retrieved from radiation spectra measured by the greenhouse gases observation satellite (GOSAT) are  
 35 shown in panels (a) and (b). Cape Grim Observatory (CGO; 41°S, 145°E) and Trinidad Head (THD;  
 36 41°N, 124°W) data are taken from the AGAGE network, NOAA global and northern hemispheric (NH)  
 37 means for  $\delta^{13}\text{C}$  are calculated from 10 and 6 sites, respectively. The PDX data adjusted to NH (period:  
 38 1977–2000) are merged with THD (period: 2001–2019) for CH<sub>4</sub> concentration and growth rate analysis,  
 39 and PDX and NOAA NH means of  $\delta^{13}\text{C}$  data are used for joint interpretation of long-term trends analysis.  
 40 The multivariate ENSO index (MEI) is shown in panel (b). Further details on data sources and processing  
 41 are available in the chapter data table (Table 5.SM.6).

42  
 43 **[END FIGURE 5.13 HERE]**

#### 44 45 46 5.2.2.2 *Anthropogenic CH<sub>4</sub> emissions*

47  
 48 The positive gradient between CH<sub>4</sub> at Cape Grim, Australia (41°S) and Trinidad Head, USA (41°N), and the  
 49 bigger difference between Trinidad Head and global mean CH<sub>4</sub> compared to that between global mean CH<sub>4</sub>  
 50 and Cape Grim, strongly suggest that the northern hemisphere is the dominant origin of anthropogenic CH<sub>4</sub>  
 51 emissions (Figure 5.13). The loss rate of CH<sub>4</sub> in troposphere does not produce a large positive north-south  
 52 hemispheric gradient in CH<sub>4</sub> due to parity in hemispheric mean OH concentration (Patra et al., 2014) or in  
 53 the case of greater OH concentrations in the northern than the southern hemisphere as simulated by the  
 54 chemistry-climate models (Naik et al., 2013). Coal mining contributed about 35% of the total CH<sub>4</sub> emissions  
 55 from all fossil fuel related sources. Top-down estimates of fossil fuel emissions (106 Tg yr<sup>-1</sup>) are smaller  
 56 than bottom-up estimates (115 Tg yr<sup>-1</sup>) during 2008–2017 (Table 5.2). Inventory-based estimates suggest that

CH<sub>4</sub> emissions from coal mining increased by 17 Tg yr<sup>-1</sup> between the periods 2002–2006 and 2008–2012, with a dominant contribution from China (Peng et al., 2016; Crippa et al., 2020; Höglund-Isaksson et al., 2020). Furthermore, top-down estimates suggest emissions from China’s coal mines have continued to grow at a slower rate after 2010 (Miller et al., 2019; Chandra et al., 2021). Emissions from oil and gas extraction and use decreased in the 1980s and 1990s, but increased in the 2000s and 2010s (Dlugokencky et al., 1994; Stern and Kaufmann, 1996; Howarth, 2019; Crippa et al., 2020). The attribution to multiple CH<sub>4</sub> source using spatially aggregated atmospheric δ<sup>13</sup>C data remained underdetermined to infer the global total emissions from the fossil fuel industry, biomass burning and agriculture (Rice et al., 2016; Schaefer et al., 2016; Schwietzke et al., 2016; Worden et al., 2017; Thompson et al., 2018).

In the agriculture and waste sectors (Table 5.2), livestock production has the largest emission source (109 Tg yr<sup>-1</sup> in 2008–2017) dominated by enteric fermentation by about 90%. Methane is formed during the storage of manure, when anoxic conditions are developed (Hristov et al., 2013). Emissions from enteric fermentation and manure have increased gradually from about 87 Tg yr<sup>-1</sup> in 1990–1999 to 109 Tg yr<sup>-1</sup> in 2008–2017 mainly due to the increase in global total animal numbers. Methane production in livestock rumens (cattle, goats, sheep, water buffalo) are affected by the type, amount and quality of feeds, energy consumption, animal size, health and growth rate, meat and milk production rate, and temperature (Broucek, 2014; Williams et al., 2015; IPCC SRCCL 5.4.3). Waste management and landfills produced 64 Tg yr<sup>-1</sup> in 2008–2017, with global emissions increasing steadily since the 1970s and despite significant declines in US, western Europe and Japan (Crippa et al., 2020; Höglund-Isaksson et al., 2020).

Emissions from rice cultivation decreased from about 45 Tg yr<sup>-1</sup> in the 1980s to about 29 Tg yr<sup>-1</sup> in the 2000–2009 but increased again slightly to 31 Tg yr<sup>-1</sup> during 2008–2017 based inventories data. However, ecosystem models showed a gradual increase with time due to climate change (*limited evidence, low agreement*) (Crippa et al., 2020; Höglund-Isaksson et al., 2020; Ito, 2020).

Biomass burning and biofuel consumption (including both natural and anthropogenic processes) caused at least 30 Tg yr<sup>-1</sup> emissions during 2008–2017 and constituted up to about 5% of global anthropogenic CH<sub>4</sub> emissions. Methane emissions from open biomass burning decreased during the past two decades mainly due to reduction of burning in savanna, grassland and shrubland (van der Werf et al., 2017; Worden et al., 2017). There is recent evidence from the tropics that fire occurrence is non-linearly related to precipitation, implying that severe droughts will increase CH<sub>4</sub> emissions from fires, particularly from the degraded peatlands (Field et al., 2016).

#### [START TABLE 5.2 HERE]

**Table 5.2: Global CH<sub>4</sub> budget.** Sources and sinks of CH<sub>4</sub> for the two recent decades from bottom-up and top-down estimations (in Tg CH<sub>4</sub> yr<sup>-1</sup>). The data are updated from (Saunois et al., 2020), for the bottom-up anthropogenic emissions (FAO, 2019; US EPA, 2019; Crippa et al., 2020; Höglund-Isaksson et al., 2020), top-down geological emissions (Schwietzke et al., 2016; Petrenko et al., 2017; Hmiel et al., 2020), and top-down sinks from 7 selected inverse models. The means (min-max) with outliers removed from both the range and the means are given. Outliers defined as >75th percentile + 3 × the interquartile range or < 25th percentile – 3 × the interquartile range. The top-down budget imbalances are calculated for each model separately and averaged. Note also the round-off error for the sources and sinks, which sometimes leads to last digit mismatch in the sums. For detailed information on datasets, see further details on data table 5.SM.6.

Tg CH <sub>4</sub> yr <sup>-1</sup>	2000–2009		2008–2017	
	Top-Down	Bottom-up	Top-Down	Bottom-up
<b>SOURCES</b>				
Natural sources	215 (176–243)	369 (245–484)	215 (183–248)	371 (245–488)
Wetlands	180 (153–196)	147 (102–178)	180 (159–199)	149 (102–182)
Other Sources	35 (21–47)	222 (143–306)	36 (21–49)	222 (143–306)

Freshwater (lakes and rivers)						159	(117-212)	
Wild animals						2	(1-3)	
Termites						9	(3-15)	
Geological (land and oceans)					23	(0-71)	45	(18-65)
Other oceanic (sea-air flux and hydrates)							6	(4-10)
Permafrost (excl. lakes and wetlands)							1	(0-1)
<b>Anthropogenic sources</b>	<b>332</b>	<b>(312-347)</b>	<b>330</b>	<b>(309-350)</b>	<b>357</b>	<b>(336-375)</b>	<b>356</b>	<b>(335-383)</b>
Agriculture & Waste	206	(198-219)	195	(185-212)	221	(209-238)	208	(192-230)
Enteric fermentation & Manure			103	(101-107)			109	(106-115)
Landfills & waste			60	(53-70)			64	(55-77)
Rice			29	(23-34)			31	(25-37)
Fossil fuels	101	(71-151)	100	(94-108)	106	(81-131)	115	(114-116)
Coal			29	(26-33)			38	(36-39)
Oil and gas			65	(60-72)			70	(68-73)
Transport			3	(1-8)			5	(1-11)
Industry			3	(0-6)			3	(1-5)
Biomass burning & biofuels	29	(23-35)	32	(24-44)	30	(22-36)	30	(22-39)
Biomass burning			19	(15-32)			17	(14-26)
Biofuels			10	(8-12)			10	(8-13)
<b>SINKS</b>								
Total chemical loss	511	(502-515)	595	(489-749)	514	(474-529)	602	(496-754)
Tropospheric OH			553	(476-677)			560	(483-682)
Stratospheric loss			31	(12-37)			31	(12-37)
Tropospheric Cl			11	(1-35)			11	(1-35)
Soil uptake	34	(27-41)	30	(11-49)	37	(27-43)	30	(11-49)
Sum of sources	548	(524-560)	699	(554-834)	576	(550-589)	727	(581-872)
Sum of sinks	546	(533-556)	625	(500-798)	551	(501-572)	632	(507-803)
Imbalance	7	(4-11)	74		21	(18-26)	95	
Atmospheric growth rate (ppb yr <sup>-1</sup> )			2 ± 4				7 ± 3	

[END TABLE 5.2 HERE]

### 5.2.2.3 Land Biospheric Emissions and Sinks

Freshwater wetlands are the single largest global natural source of CH<sub>4</sub> into the atmosphere, accounting for about 26% of the total CH<sub>4</sub> source (*robust evidence, medium agreement*). Progress has been made since AR5 (Ciais et al., 2013) in better constraining freshwater lake and river emissions and reducing double counting with wetland emissions. Bottom-up and top-down estimates for 2008–2017 are 149 and 180 Tg yr<sup>-1</sup>, respectively, with a top-down uncertainty range of 159–199 Tg yr<sup>-1</sup> (Table 5.2). The large uncertainties stem from challenges in mapping wetland area and temporal dynamics, and in scaling methane production, transport and consumption processes, that are measured with small chambers or flux towers, to landscape estimates (Pham-Duc et al., 2017). Both the top-down and bottom-up estimates presented in Table 5.2 indicate little increase in wetland CH<sub>4</sub> emissions during the last three decades, with the new estimates being slightly smaller than in AR5 due to updated wetland maps and ecosystem model simulations (Melton et al., 2013; Poulter et al., 2017). The wetland emissions show strong interannual variability due to the changes in inundated land area, air temperature and microbial activity (Bridgham et al., 2013). Present terrestrial ecosystem model simulated CH<sub>4</sub> emission variability does not produce strong correlation with the ENSO cycle (Cross-Chapter Box 5.2, Figure 2), although observation evidence is emerging for lower CH<sub>4</sub> emissions during El Niños and greater emissions during the La Niña (Pandey et al., 2017).

Trees in upland and wetland forests contribute to CH<sub>4</sub> emissions by abiotic production in the canopy, by the methanogenesis taking place in the stem, and by conducting CH<sub>4</sub> from soil into the atmosphere (Covey and

1 Megonigal, 2019). There is emerging evidence of the important role of trees in transporting and conducting  
2 CH<sub>4</sub> from soils into the atmosphere especially in tropics (Pangala et al., 2017), whereas direct production of  
3 CH<sub>4</sub> by vegetation only has a minor contribution (*limited evidence, high agreement*) (Bruhn et al., 2012;  
4 Covey and Megonigal, 2019). The contribution of trees in transporting CH<sub>4</sub> may further widen the gap  
5 between the bottom-up and top-down estimates in the global budget, particularly needing a reassessment of  
6 emissions in the tropics and in forested wetlands of temperate and boreal regions (Pangala et al., 2017;  
7 Jeffrey et al., 2019; Welch et al., 2019; Sjögersten et al., 2020).

8  
9 Microbial methane uptake by soil comprises up to 5% (30 Tg yr<sup>-1</sup>) of the total CH<sub>4</sub> sink in 2008–2017 (Table  
10 5.2). There is evidence from experimental and modelling studies of increasing soil microbial uptake due to  
11 increasing temperature (Yu et al., 2017), although evidence for decreasing CH<sub>4</sub> consumption, possibly linked  
12 to precipitation changes, also exist (Ni and Groffman, 2018). The estimate of global methane loss by  
13 microbial oxidation in upland soils has been lowered marginally by 4 Tg yr<sup>-1</sup>, compared to 34 Tg yr<sup>-1</sup> in  
14 AR5, for the period 2000–2009. Termites, an infraorder of insects (Isoptera) found in almost all landmasses,  
15 emitted about 9 Tg yr<sup>-1</sup> of CH<sub>4</sub> in 2000–2009, and increased emissions from the insects and other arthropods  
16 are projected (Brune, 2018).

#### 19 5.2.2.4 Ocean and Inland Water Emissions and Sinks

21 In AR5, the ocean CH<sub>4</sub> emissions were reported together with geological emissions summing up to 54 (33–  
22 75) Tg yr<sup>-1</sup>. Coastal oceans, fjords and mud volcanos are major source of CH<sub>4</sub> in the marine environment, but  
23 CH<sub>4</sub> flux measurements are sparse. Saunio et al. (2020) estimate that the oceanic budget, including  
24 biogenic, geological and hydrate emissions from coastal and open ocean, is 6 (range 4–10) Tg yr<sup>-1</sup> for the  
25 2000s, which is in good agreement with an air-sea flux measurement-based estimate of 6–12 Tg yr<sup>-1</sup> (Weber  
26 et al., 2019). When estuaries are included, the total oceanic budget is 9–22 Tg yr<sup>-1</sup>, with a mean value of 13  
27 Tg yr<sup>-1</sup>. A recent synthesis suggests that CH<sub>4</sub> emissions from shallow coastal ecosystems, particularly from  
28 mangroves, can be as high as 5–6 Tg yr<sup>-1</sup> (Al-Haj and Fulweiler, 2020). The reservoir emissions, including  
29 coastal wetlands and tidal flats, contribute up to 13 Tg yr<sup>-1</sup> (Borges and Abril, 2011; Deemer et al., 2016).  
30 Methane seepage from the Arctic shelf, possibly triggered by the loss of geological storage due to warming  
31 and thawing of permafrost and hydrate decomposition, has a wide estimated range of 0.0–17 Tg yr<sup>-1</sup>  
32 (Shakhova et al., 2010, 2014, 2017; Berchet et al., 2016); advanced eddy covariance measurements put the  
33 best estimate at just about 3 Tg yr<sup>-1</sup> from the East Siberian Arctic shelf (Thornton et al., 2020). The current  
34 flux is expected to be a mix of pre-industrial and climate change-driven fluxes, CH<sub>4</sub> seepage is anticipated to  
35 increase in a warmer world (Dean et al., 2018).

37 All geological sources around the world, including the coastal oceans and fjords, are estimated to emit CH<sub>4</sub>  
38 in the range of 35–76 Tg yr<sup>-1</sup> (Etiope et al., 2019). There is evidence that the ventilation of geological CH<sub>4</sub> is  
39 *likely* to be smaller than 15 Tg yr<sup>-1</sup> (Petrenko et al., 2017; Hmiel et al., 2020). A lower geological CH<sub>4</sub>  
40 ventilation will reduce the gap between bottom-up estimations and that are used in top-down models (Table  
41 5.2), but widen the gap in the ratio of fossil-fuel derived sources to the biogenic sources for matching the  
42 Δ<sup>14</sup>C-CH<sub>4</sub> observations.

44 Inland water (lakes, rivers, streams, ponds, estuaries) emissions are proportionally the largest source of  
45 uncertainty in the CH<sub>4</sub> budget. Since AR5 (Ciais et al., 2013), the inland water CH<sub>4</sub> source has been revised  
46 from 8–73 Tg yr<sup>-1</sup> (1980s) to 117–212 Tg yr<sup>-1</sup> (2000s) with the availability of more observational data and  
47 improved areal estimates (Bastviken et al., 2011; Deemer et al., 2016; Stanley et al., 2016; DelSontro et al.,  
48 2018; Saunio et al., 2020). A large spatial and temporal variation in lake and river CH<sub>4</sub> fluxes (Wik et al.,  
49 2016; Crawford et al., 2017; Natchimuthu et al., 2017) and uncertainties in the global area of them (Allen  
50 and Pavelsky, 2018), together with a relatively small number of observations, varying measurement  
51 methods, for example those neglecting ebullition, varying upscaling methods and lack of appropriate process  
52 make the bottom-up CH<sub>4</sub> emission estimate uncertain (Sanchez et al., 2019; Engram et al., 2020; Zhang et  
53 al., 2020a). Accordingly, there is no clear accounting of inland waters in top-down budgets, which is the  
54 main reason for the large gap in bottom-up and top-down estimates of “other sources” in the CH<sub>4</sub> budget  
55 (Table 5.2). Despite recent progress in separating wetlands from inland waters, there is double-counting in

1 the bottom-up estimates of their emissions (Thornton et al., 2016). Although there is evidence that regional  
2 human activities and warming both increase inland water CH<sub>4</sub> emissions (Beaulieu et al., 2019), the increase  
3 in the decadal emissions since AR5 (Ciais et al., 2013) rather reflect improvements in the estimate (*medium*  
4 *confidence*), due to updates in the datasets and new upscaling approaches (Saunois et al., 2020).

#### 5.2.2.5 CH<sub>4</sub> Budget

9 A summary of top-down and bottom-up estimations of CH<sub>4</sub> emissions and sinks for the period 2008–2017 is  
10 presented in Figure 5.14 (details in Table 5.2 and the associated text for the emissions). In addition to 483-  
11 682 Tg yr<sup>-1</sup> loss of CH<sub>4</sub> in the troposphere by reaction with OH, 1–35 Tg yr<sup>-1</sup> of CH<sub>4</sub> loss is estimated to  
12 occur in the lower troposphere due to Cl but are not included in the top-down models as shown in Table 5.2  
13 (Hossaini et al., 2016; Gromov et al., 2018; Wang et al., 2019b). The decadal mean CH<sub>4</sub> burden/imbalance  
14 have increased at the rate of 30, 12, 7 and 21 Tg yr<sup>-1</sup> in the 1980s (1980–1989), 1990s (1990–1999), 2000s  
15 (2000–2009) and the most recent decade (2008–2017), respectively (*virtually certain*), as can be estimated  
16 from observed atmospheric growth rate (Cross-Chapter Box 5.2, Figure 1).

18 Recent analysis using Δ<sup>14</sup>C-CH<sub>4</sub> in ice samples suggest that CH<sub>4</sub> emissions from fossil fuels exploitation are  
19 responsible for 30% of total CH<sub>4</sub> emissions (Lassey et al., 2007; Hmiel et al., 2020), which is largely  
20 inconsistent with sectorial budgets where fossil fuel emissions add up to 20% only (Ciais et al., 2013).  
21 However, recent model simulations produce fairly consistent δ<sup>13</sup>C-CH<sub>4</sub> values and trends as observed in the  
22 atmospheric samples using 20% fossil fuel emission fraction (Ghosh et al., 2015; Warwick et al., 2016;  
23 Fujita et al., 2020; Strode et al., 2020). Further research is needed in order to clarify relative roles of CH<sub>4</sub>  
24 emissions from fossil fuel exploitation and freshwater components. A key challenge is to accommodate  
25 higher estimated emissions from these two components without a major increase in the sinks for explaining  
26 the carbon and hydrogen isotopes variabilities at the same time.

[START FIGURE 5.14 HERE]

31 **Figure 5.14: Global methane (CH<sub>4</sub>) budget (2008–2017).** Values and data sources as in Table 5.2 (in TgCH<sub>4</sub>). The  
32 atmospheric stock is calculated from mean CH<sub>4</sub> concentration, multiplying a factor of 2.75 ± 0.015 Tg  
33 ppb<sup>-1</sup>, which accounts for the uncertainties in global mean CH<sub>4</sub> (Chandra et al., 2021). Further details on  
34 data sources and processing are available in the chapter data table (Table 5.SM.6).

[END FIGURE 5.14 HERE]

[START CROSS-CHAPTER BOX 5.2]

#### Cross-Chapter Box 5.2: Drivers of atmospheric methane changes during 1980–2019

43 **Contributors:** Prabir K. Patra (Japan/India), Josep G. Canadell (Australia), Frank Dentener (EU,  
44 Netherlands), Xin Lan (USA), Vaishali Naik (USA)

46 The atmospheric methane (CH<sub>4</sub>) growth rate has varied widely over the past three decades, and the causes of  
47 which have been extensively studied since AR5. The mean growth rate decreased from 15 ± 5 ppb yr<sup>-1</sup> in the  
48 1980s to 0.48 ± 3.2 ppb yr<sup>-1</sup> during 2000–2006 (the so-called quasi-equilibrium phase) and returned to an  
49 average rate of 7.6 ± 2.7 ppb yr<sup>-1</sup> in the past decade (2010–2019) (based on data in Figure 5.14).

50 Atmospheric CH<sub>4</sub> grew faster (9.3 ± 2.4 ppb yr<sup>-1</sup>) over the last six years (2014–2019) – a period with  
51 prolonged El Niño conditions, which contributed to high CH<sub>4</sub> growth rates consistent with behaviour during  
52 previous El Niño events (Figure 5.14b). Because of large uncertainties in both the emissions and sinks of  
53 CH<sub>4</sub>, it has been challenging to quantify accurately the methane budget and ascribe reasons for the growth  
54 over 1980–2019. In the context of CH<sub>4</sub> emissions mitigation, it is critical to understand if the changes in  
55 growth rates are caused by emissions from human activities or by natural processes responding to changing

1 climate. If CH<sub>4</sub> continues to grow at rates similar to those observed over the past decade, it will contribute to  
 2 decadal scale climate change and hinder the achievement of the long-term temperature goals of the Paris  
 3 Agreement (Nisbet et al., 2019)(7.3.2.2).

4  
 5 Cross-Chapter Box 5.2 Figure 1 shows the decadal CH<sub>4</sub> budget derived from the Global Carbon Project  
 6 (GCP)-CH<sub>4</sub> synthesis for 1980s, 1990s and 2000s (Kirschke et al., 2013), and for 2010–2017 (Saunois et al.,  
 7 2020). The imbalance of the sources and sinks estimated by atmospheric inversions (red bars) can be used to  
 8 explain the changes in CH<sub>4</sub> concentration increase rates between the decades (Table 5.2).

9  
 10  
 11 **[START CROSS-CHAPTER BOX 5.2, FIGURE 1 HERE]**

12  
 13 **Cross-Chapter Box 5.2, Figure 1: Methane sources and sinks for four decades from atmospheric inversions with**  
 14 **the budget imbalance** (source-sink; red bars) (plotted on the left y-axis). Top-down  
 15 analysis from (Kirschke et al., 2013; Saunois et al., 2020), The global CH<sub>4</sub>  
 16 concentration seen in the black line (plotted on the right y-axis), representing NOAA  
 17 observed global monthly mean atmospheric CH<sub>4</sub> in dry-air mole fractions for 1983–  
 18 2019 (Chapter 2, Annex III. Natural sources include emissions from natural  
 19 wetlands, lakes and rivers, geological sources, wild animals, termites, wildfires,  
 20 permafrost soils, and oceans. Anthropogenic sources include emissions from enteric  
 21 fermentation and manure, landfills, waste and wastewater, rice cultivation, coal  
 22 mining, oil and gas industry, biomass and biofuel burning. The top-down total sink  
 23 is determined from global mass balance includes chemical losses due to reactions  
 24 with hydroxyl (OH), atomic chlorine (Cl), and excited atomic oxygen (O<sup>1</sup>D), and  
 25 oxidation by bacteria in aerobic soils (Table 5.2). Further details on data sources and  
 26 processing are available in the chapter data table (Table 5.SM.6).

27  
 28 **[END CROSS-CHAPTER BOX 5.2, FIGURE 1 HERE]**

29  
 30  
 31 Since AR5, many studies have discussed the role of different source categories in explaining the increase in  
 32 CH<sub>4</sub> growth rate since 2007 and a coincident decrease of δ<sup>13</sup>C–CH<sub>4</sub> and δD–CH<sub>4</sub> isotopes (ref. Figure 5.13;  
 33 Rice et al., 2016). Both <sup>13</sup>C and D are enriched in mass-weighted average source signatures for CH<sub>4</sub>  
 34 emissions from thermogenic sources (e.g. coal mining, oil and gas industry) and pyrogenic (biomass  
 35 burning) sources, and depleted in biogenic (e.g. wetlands, rice paddies, enteric fermentation, landfill and  
 36 waste) sources. Proposed hypotheses for CH<sub>4</sub> growth (2007–2017) are inconclusive and vary from a  
 37 concurrent decrease in thermogenic and increase in wetland and other biogenic emissions (Nisbet et al.,  
 38 2016; Schwietzke et al., 2016), increase in from emissions agriculture in the tropics (Schaefer et al., 2016), a  
 39 concurrent reduction in pyrogenic and increase in thermogenic emissions (Worden et al., 2017), or emission  
 40 increase from biogenic sources and a slower increase in emissions from thermogenic sources compared to  
 41 inventory emissions (Patra et al., 2016; Thompson et al., 2018; Jackson et al., 2020; Chandra et al., 2021).

42  
 43 A few studies have emphasised the role of chemical destruction by OH, the primary sink of methane, in  
 44 driving changes in the growth of atmospheric methane abundance, in particular after 2006 (Rigby et al.,  
 45 2017; Turner et al., 2017). Studies applying three-dimensional atmospheric inversion (McNorton et al.,  
 46 2018), simple multi-species inversion (Thompson et al., 2018), as well as empirical method using a variety  
 47 of observational constraints based on OH chemistry (Nicely et al., 2018; Patra et al., 2021), do not find  
 48 trends in OH large enough to explain the methane changes post-2006. On the contrary, global chemistry-  
 49 climate models based on fundamental principles of atmospheric chemistry and known emission trends of  
 50 anthropogenic non-methane SLCFs simulate an increase in OH over this period (Zhao et al., 2019;  
 51 Stevenson et al., 2020) (see Section 6.2.3). These contrasting lines of evidence suggest that OH changes may  
 52 have had a small moderating influence on methane growth since 2007 (*low confidence*).

53  
 54 Cross-Chapter Box 5.2 Figure 2 shows that modelled wetland emission anomalies for all regions did not  
 55 exhibit statistically significant trends (*high agreement between models, medium evidence*). Thus, the inter-  
 56 decadal difference of total CH<sub>4</sub> emissions derived from inversion models and wetland emissions, arises

1 mainly from anthropogenic activities. The timeseries of regional emissions suggest that progress towards  
 2 atmospheric CH<sub>4</sub> quasi-equilibrium was primarily driven by reductions in anthropogenic (fossil fuel  
 3 exploitation) emissions in Europe, Russia and temperate North America over 1988–2000. In the global  
 4 totals, emissions equalled loss in the early 2000s. The growth since 2007 is driven by increasing agricultural  
 5 emissions from East Asia (1997–2017), West Asia (2005–2017), Brazil (1988–2017) and Northern Africa  
 6 (2005–2017), and fossil fuel exploitations in temperate North America (2010–2017) (Lan et al., 2019; Crippa  
 7 et al., 2020; Höglund-Isaksson et al., 2020; Jackson et al., 2020; Chandra et al., 2021).

8  
 9  
 10 **[START CROSS-CHAPTER BOX 5.2, FIGURE 2 HERE]**

11  
 12 **Cross-Chapter Box 5.2, Figure 2: Anomalies in global and regional methane (CH<sub>4</sub>) emissions for 1988–2017.** Map  
 13 in the centre shows mean CH<sub>4</sub> emission for 2010–2016. Multi-model mean (line)  
 14 and 1-σ standard deviations (shaded) for 2000–2017 are shown for 9 surface CH<sub>4</sub>  
 15 and 10 satellite XCH<sub>4</sub> inversions, and 22 wetland models or model variants that  
 16 participated in GCP-CH<sub>4</sub> budget assessment (Saunois et al., 2020). The results for  
 17 the period before 2000 are available from two inversions, 1) using 19 sites (Chandra  
 18 et al., 2021; also used for the 2010–2016 mean emission map) and for global totals  
 19 (Bousquet et al., 2006). The long-term mean values for 2010–2016 (common for all  
 20 GCP-CH<sub>4</sub> inversions), as indicated within each panel separately, is subtracted from  
 21 the annual-mean time series for the calculation of anomalies for each region. Further  
 22 details on data sources and processing are available in the chapter data table (Table  
 23 5.SM.6).

24  
 25 **[END CROSS-CHAPTER BOX 5.2, FIGURE 2 HERE]**

26  
 27  
 28 Evidence from emission inventories at country level and regional scale inverse modelling that CH<sub>4</sub> growth  
 29 rate variability during the 1988 through 2017 is closely linked to anthropogenic activities (*medium*  
 30 *agreement*). Isotopic composition observations and inventory data suggest that concurrent emission changes  
 31 from both fossil fuels and agriculture are playing roles in the resumed CH<sub>4</sub> growth since 2007 (*high*  
 32 *confidence*). Shorter-term decadal variability is predominantly driven by the influence of El Niño Southern  
 33 Oscillation on emissions from wetlands and biomass burning (Cross-Chapter Box 5.2 Figure 2), and loss due  
 34 to OH variations (*medium confidence*), but lacking quantitative contribution from each of the sectors. By  
 35 synthesising all available information regionally from a-priori (bottom-up) emissions, satellite and surface  
 36 observations, including isotopic information, and inverse modelling (top-down), the capacity to track and  
 37 explain changes in and drivers of natural and anthropogenic CH<sub>4</sub> regional and global emissions has been  
 38 improved since the AR5, but fundamental uncertainties related to OH variations remain unchanged.

39  
 40 **[END CROSS-CHAPTER BOX 5.2]**

41  
 42  
 43 **5.2.3 N<sub>2</sub>O: Trends, Variability and Budget**

44  
 45 In natural ecosystems, nitrous oxide (N<sub>2</sub>O) is primarily produced as a by-product during the remineralisation  
 46 of organic matter via the primary processes of nitrification and denitrification (Butterbach-Bahl et al., 2013;  
 47 Voss et al., 2013). The net N<sub>2</sub>O production is highly sensitive to local environmental conditions such as  
 48 temperature, oxygen concentrations, pH and the concentrations of ammonium and nitrate, amongst others,  
 49 causing strong variability of N<sub>2</sub>O emissions in time and space even at small scales. Changes in the  
 50 atmospheric abundance of N<sub>2</sub>O result largely from the balance of the net N<sub>2</sub>O sources on land and ocean, and  
 51 the photochemical destruction of N<sub>2</sub>O in the stratosphere.

52  
 53 Since AR5 (WGI, 6.4.3), improved understanding of N<sub>2</sub>O sources allows for a more comprehensive  
 54 assessment of the global N<sub>2</sub>O budget (Table 5.3). This progress is based on extended atmospheric  
 55 observations (Francey et al., 2003; Elkins et al., 2018; Prinn et al., 2018), improved atmospheric N<sub>2</sub>O  
 56 inversions (Saikawa et al., 2014; Thompson et al., 2019), updated and expanded inventories of N<sub>2</sub>O sources



1 (Janssens-Maenhout et al., 2017; Winiwarter et al., 2018), as well as improved bottom-up estimate of  
 2 freshwater, ocean and terrestrial sources (Martinez-Rey et al., 2015; Landolfi et al., 2017; Buitenhuis et al.,  
 3 2018a; Lauerwald et al., 2019; Maavara et al., 2019; Tian et al., 2019).

4  
 5 The human perturbation of the natural nitrogen cycle through the use of synthetic fertilisers and manure, as  
 6 well as nitrogen deposition resulting from land-based agriculture and fossil fuel burning has been the largest  
 7 driver of the increase in atmospheric N<sub>2</sub>O of  $31.0 \pm 0.5$  parts per billion (ppb) (10%) between 1980 and 2019  
 8 (*robust evidence, high agreement*) (Tian et al., 2020). The long atmospheric lifetime of N<sub>2</sub>O implies that it  
 9 will take more than a century before atmospheric abundances stabilise after the stabilisation of global  
 10 emissions. The rise of atmospheric N<sub>2</sub>O is of concern, not only because of its contribution to the  
 11 anthropogenic radiative forcing (see Chapter 7), but also because of the importance of N<sub>2</sub>O in stratospheric  
 12 ozone loss (Ravishankara et al., 2009; Fleming et al., 2011; Wang et al., 2014a).

### 13 14 15 5.2.3.1 Atmosphere

16  
 17 The tropospheric abundance of N<sub>2</sub>O was  $332.1 \pm 0.4$  ppb in 2019 (Figure 5.15), which is 23% higher than  
 18 pre-industrial levels of  $270.1 \pm 6.0$  ppb (*robust evidence, high agreement*). Current estimates are based on  
 19 atmospheric measurements with high accuracy and density (Francey et al., 2003; Elkins et al., 2018; Prinn et  
 20 al., 2018), and pre-industrial estimates are based on multiple ice-core records (see Section 2.2.3.2.3). The  
 21 average annual tropospheric growth rate was  $0.85 \pm 0.03$  ppb yr<sup>-1</sup> during the period 1995 to 2019 (Figure  
 22 5.15a). The atmospheric growth rate increased by about 20% between the decade of 2000 to 2009 and the  
 23 most recent decade of 2010 to 2019 ( $0.95 \pm 0.04$  ppb yr<sup>-1</sup>) (*robust evidence, high agreement*). The growth  
 24 rate in 2010–2019 was also higher than during 1970–2000 ( $0.6–0.8$  ppb yr<sup>-1</sup> (Ishijima et al., 2007)) and the  
 25 thirty-year period prior to 2011 ( $0.73 \pm 0.03$  ppb yr<sup>-1</sup>), as reported by AR5. New evidence since AR5 (WGI,  
 26 6.4.3) confirms that in the tropics and sub-tropics, large inter-annual variations in the atmospheric growth  
 27 rate are negatively correlated with the multivariate ENSO index (MEI) and associated anomalies in land and  
 28 ocean fluxes (Ji et al., 2019; Thompson et al., 2019; Yang et al., 2020c) (Figure 5.15a).

29  
 30 As assessed by SRCCL (IPCC, 2019a), combined firn, ice, air and atmospheric measurements show that the  
 31 <sup>15</sup>N/<sup>14</sup>N isotope ratio (*robust evidence, high agreement*) as well as the predominant position of the <sup>15</sup>N atom  
 32 in atmospheric N<sub>2</sub>O (*limited evidence, low agreement*) in N<sub>2</sub>O has changed since 1940 (Figure 5.15b, c)  
 33 whereas they were relatively constant in the pre-industrial period (Ishijima et al., 2007; Park et al., 2012;  
 34 Prokopiou et al., 2017, 2018). SRCCL concluded that this change indicates a shift in the nitrogen-substrate  
 35 available for de-nitrification, and the relative contribution of nitrification to the global N<sub>2</sub>O source (*robust*  
 36 *evidence, high agreement*), which are associated with increased fertiliser use in agriculture (Park et al., 2012;  
 37 Snider et al., 2015; Prokopiou et al., 2018).

38  
 39 Since AR5 (WGI, 6.4.3), the mean atmospheric lifetime of N<sub>2</sub>O has been revised to  $116 \pm 9$  years (Prather et  
 40 al., 2015). The small negative feedback of the N<sub>2</sub>O lifetime to increasing atmospheric N<sub>2</sub>O results in a  
 41 slightly lower residence time ( $109 \pm 10$  years) of N<sub>2</sub>O perturbations compared with that assessed by AR5  
 42 (118–131 years) (Prather et al., 2015). The dominant N<sub>2</sub>O loss occurs through photolysis and oxidation by  
 43 O(1D) radicals in the Stratosphere and amounts to approximately 13.1 (12.4–13.6) TgN yr<sup>-1</sup> (Minschwaner et  
 44 al., 1993; Prather et al., 2015; Tian et al., 2020).

45  
 46  
 47 **[START FIGURE 5.15 HERE]**

#### 48 49 **Figure 5.15: Changes in atmospheric nitrous oxide (N<sub>2</sub>O) and its isotopic composition since 1940. (a)**

50 Atmospheric N<sub>2</sub>O abundance (parts per billion, ppb) and growth rate (ppb yr<sup>-1</sup>), (b) δ<sup>15</sup>N of atmospheric  
 51 N<sub>2</sub>O, and (c) alpha-site <sup>15</sup>N–N<sub>2</sub>O. Estimate are based on direct atmospheric measurements in the AGAGE  
 52 , CSIRO, and NOAA networks (Prinn et al., 2000, 2018; Francey et al., 2003; Hall et al., 2007; Elkins et  
 53 al., 2018), archived air samples from Cape Grim, Australia (Park et al., 2012), and firn air from NGRIP  
 54 Greenland and H72 Antarctica (Ishijima et al., 2007), Law Dome Antarctica (Park et al., 2012), as well as  
 55 a collection of firn ice samples from Greenland (Prokopiou et al., 2017, 2018). Shading in (a) is based on  
 56 the multivariate ENSO index, with red indicating El Niño conditions (Wolter and Timlin, 1998). Further

1 details on data sources and processing are available in the chapter data table (Table 5.SM.6).

2  
3 **[END FIGURE 5.15 HERE]**  
4  
5

### 6 5.2.3.2 Anthropogenic N<sub>2</sub>O Emissions

7  
8 AR5 (WGI, 6.4.3) and SRCCL (2.3.3) concluded that agriculture is the largest anthropogenic source of N<sub>2</sub>O  
9 emissions. Since SRCCL (2.3.3), a new synthesis of inventory-based and modelling studies shows that the  
10 widespread use of synthetic fertilisers and manure on cropland and pasture, manure management and  
11 aquaculture resulted in 3.8 (2.5–5.8) TgN yr<sup>-1</sup> (average 2007–2016) (*robust evidence, high agreement*)  
12 (Table 5.3) (Winiwarter et al., 2018; FAO, 2019; Janssens-Maenhout et al., 2019; Tian et al., 2020).  
13 Observations from field-measurements (Song et al., 2018), inventories (Wang et al., 2020) and atmospheric  
14 inversions (Thompson et al., 2019) further corroborate the assessment of the SRCCL that there is a non-  
15 linear relationship between N<sub>2</sub>O emissions and nitrogen input, implying an increasing fraction of fertiliser  
16 lost as N<sub>2</sub>O with larger fertiliser excess (*medium evidence, high agreement*). Several studies using  
17 complementary methods indicate that agricultural N<sub>2</sub>O emissions have increased by more than 45% since the  
18 1980s (*high confidence*) (Davidson, 2009; Janssens-Maenhout et al., 2017; Winiwarter et al., 2018; Tian et  
19 al., 2020) (Figure 5.16, Table 5.3), mainly due to the increased use of nitrogen fertiliser and manure. N<sub>2</sub>O  
20 emissions from aquaculture are amongst the fastest rising contributors of N<sub>2</sub>O emissions, but their overall  
21 magnitude is still small in the overall N<sub>2</sub>O budget (Tian et al., 2020).  
22

23  
24 **[START FIGURE 5.16 HERE]**  
25

26 **Figure 5.16: Decadal mean nitrous oxide (N<sub>2</sub>O) emissions for 2007–2016 and its change since 1850 based on**  
27 **process-model projections.** The total effect including that from anthropogenic nitrogen additions  
28 (atmospheric deposition, manure addition, fertiliser use and land-use) is evaluated against the background  
29 flux driven by changes in atmospheric CO<sub>2</sub> concentration, and climate change. Fluxes are derived from the  
30 N<sub>2</sub>O Model Intercomparison Project ensemble of terrestrial biosphere models (Tian et al., 2019) and three  
31 ocean biogeochemical models (Landolfi et al., 2017; Battaglia and Joos, 2018a; Buitenhuis et al., 2018b).  
32 Further details on data sources and processing are available in the chapter data table (Table 5.SM.6).  
33

34 **[END FIGURE 5.16 HERE]**  
35  
36

37 The principal non-agricultural anthropogenic sources of N<sub>2</sub>O are industry, specifically chemical processing,  
38 wastewater, and the combustion of fossil fuels (Table 5.3). Industrial emissions of N<sub>2</sub>O mainly due to nitric  
39 and adipic acid production have decreased in North America and Europe since the wide-spread installation  
40 of abatement technologies in the 1990s (Pérez-Ramírez et al., 2003; Lee et al., 2011; Janssens-Maenhout et  
41 al., 2019). There is still considerable uncertainty in industrial emissions from other regions of the world with  
42 contrasting trends between inventories (Thompson et al., 2019). Globally, industrial emissions and emissions  
43 from fossil fuel combustion by stationary sources, such as power plants, as well as smaller emissions from  
44 mobile sources (e.g. road transport and aviation) have remained nearly constant between 1980s and 2007-  
45 2016 (*moderate evidence, medium agreement*) (Janssens-Maenhout et al., 2017; Winiwarter et al., 2018;  
46 Tian et al., 2020). Wastewater N<sub>2</sub>O emissions, including those from domestic and industrial sources have  
47 increased from 0.2 (0.1–0.3) TgN yr<sup>-1</sup> to 0.35 (0.2–0.5) TgN yr<sup>-1</sup> between the 1980s and 2007–2016 (Tian et  
48 al., 2020).  
49

50 Biomass burning from crop residue burning, grassland, savannah and forest fires, as well as biomass burnt in  
51 household stoves, releases N<sub>2</sub>O during the combustion of organic matter. Updated inventories since AR5  
52 (WGI, 6.4.3) result in a lower range of the decadal mean emissions of 0.6 (0.5–0.8) TgN yr<sup>-1</sup> (van der Werf et  
53 al., 2017; Tian et al., 2020). The attribution of grassland, savannah or forest fires to natural or anthropogenic  
54 origins is uncertain, preventing a separation of the biomass burning source into natural and anthropogenic.  
55  
56

### 5.2.3.3 Emissions from Ocean, Inland Water Bodies and Estuaries

Since AR5 (WGI, 6.4.3), new estimates of the global ocean N<sub>2</sub>O source derived from ocean biogeochemistry models are 3.4 (2.5–4.3) TgN yr<sup>-1</sup> for the period 2007–2016 (Manizza et al., 2012; Suntharalingam et al., 2012; Martinez-Rey et al., 2015; Landolfi et al., 2017; Buitenhuis et al., 2018a; Tian et al., 2020) (Figure 5.16). This is slightly lower than climatological estimates from empirically based-methods and surface ocean data syntheses (Bianchi et al., 2012; Yang et al., 2020c). Nitrous oxide processes in coastal upwelling zones continue to be poorly represented in global estimates of marine N<sub>2</sub>O emissions (Kock et al., 2016), but may account for an additional 0.2 to 0.6 TgN yr<sup>-1</sup> of the global ocean source (Seitzinger et al., 2000; Nevison et al., 2004).

In the oxic ocean (> 97% of ocean volume), nitrification is believed to be the primary N<sub>2</sub>O source (Freing et al., 2012). In sub-oxic ocean zones (see Section 5.3), where denitrification prevails, higher N<sub>2</sub>O yields and turnover rates make these regions potentially significant sources of N<sub>2</sub>O (Arévalo-Martínez et al., 2015; Babbin et al., 2015; Ji et al., 2015). The relative proportion of ocean N<sub>2</sub>O from oxygen-minimum zones is highly uncertain (Zamora et al., 2012). Estimates derived from in situ sampling, particularly in the eastern tropical Pacific, suggest significant fluxes from these regions, and potentially accounting for up to 50% of the global ocean source (Codispoti, 2010; Arévalo-Martínez et al., 2015; Babbin et al., 2015). However, recent global-scale analyses estimate lower contributions (4 to 7%, Battaglia and Joos, 2018; Buitenhuis et al., 2018). Further investigation is required to reconcile these estimates and provide improved constraints on the N<sub>2</sub>O source from low-oxygen zones.

Atmospheric deposition of anthropogenic N on oceans can stimulate marine productivity and influence ocean emissions of N<sub>2</sub>O. New ocean model analyses since AR5 (WGI, 6.4.3), suggest a relatively modest global potential impact of 0.01–0.32 TgN yr<sup>-1</sup> (pre-industrial to present-day) equivalent to 0.5–3.3% of the global ocean N<sub>2</sub>O source (Suntharalingam et al., 2012; Jickells et al., 2017; Landolfi et al., 2017). However, larger proportionate impacts are predicted in nitrogen-limited coastal and inland waters down-wind of continental pollution outflow, such as the Northern Indian Ocean (Jickells et al., 2017; Suntharalingam et al., 2019).

Inland waters and estuaries are generally sources of N<sub>2</sub>O as a result of nitrification and denitrification of dissolved inorganic nitrogen, however, they can serve as N<sub>2</sub>O sinks in specific conditions (Webb et al., 2019). Since AR5 (WGI, 6.4.3), improved emission factors including their spatio-temporal scaling, and consideration of transport within the aquatic system allow to better constrain these emissions (Murray et al., 2015; Hu et al., 2016; Lauerwald et al., 2019; Maavara et al., 2019; Kortelainen et al., 2020; Yao et al., 2020). Despite uncertainties because of the side-effects of canals and reservoirs on nutrient cycling, these advances permit to attribute a fraction of inland water N<sub>2</sub>O emissions to anthropogenic sources (Tian et al., 2020), which contributes to the increased anthropogenic share of the global N<sub>2</sub>O source in this report compared to AR5 (Ciais et al., 2013). As indirect consequence of agricultural nitrogen-use and waste-water treatment, the anthropogenic emissions from inland waters have increased by about a quarter (0.1 TgN yr<sup>-1</sup>) between the 1980s and 2007–2016 (Tian et al., 2020).

### 5.2.3.4 Emissions and Sinks in Non-Agricultural Land

Soils are the largest natural source of N<sub>2</sub>O, arising primarily from nitrogen processing associated with microbial nitrification and denitrification (Butterbach-Bahl et al., 2013; Snider et al., 2015) (Table 5.3). Under some conditions, soils can also act as a net sink of N<sub>2</sub>O, but this effect is small compared to the overall source (Schlesinger, 2013). Since AR5 (WGI, 6.4.3), improved global process-based models (Tian et al., 2019) suggest a present-day source of 6.7 (5.3–8.1) TgN yr<sup>-1</sup> (2007–2016 average), which is consistent with the estimate in AR5. Process-based models and inventory-based methods show that increased N deposition has enhanced terrestrial N<sub>2</sub>O emissions by 0.8 (0.4–1.4 TgN yr<sup>-1</sup>) relative to approximately pre-industrial times, and by 0.2 (0.1–0.2) TgN yr<sup>-1</sup> between the 1980s and 2007–2016 (*limited evidence, medium agreement*) (Figure 5.16) (Tian et al., 2019). This estimate is at the high end of the range reported in AR5 (WGI, 6.4.3). Model projections further show that global warming has led to increased soil N<sub>2</sub>O emissions of

1 0.8 (0.3–1.3) TgN yr<sup>-1</sup> since approximately pre-industrial times, of which about half occurred since the 1980s  
2 (*limited evidence, high agreement*) (Tian et al., 2019, 2020).

3  
4 SRCCL assessed that deforestation and other forms of land-use change significantly alter terrestrial N<sub>2</sub>O  
5 emissions through emission pulses following conversions, resulting generally in long-term reduced  
6 emissions in unfertilised ecosystems (*medium evidence, high agreement*). This conclusion is supported by a  
7 recent study demonstrating that the deforestation-pulse effect is offset by the effect of reduced area of mature  
8 tropical forests (Tian et al., 2020).

9  
10 Uncertainties remain in process-based models with respect to their ability to capture the complicated  
11 responses of terrestrial N<sub>2</sub>O emissions to rain pulses, freeze-thaw cycles and the net consequences of  
12 elevated levels of CO<sub>2</sub> accurately (Tian et al., 2019). Emerging literature suggests that permafrost thaw may  
13 contribute significantly to arctic N<sub>2</sub>O emissions (Voigt et al., 2020), but these processes are not yet  
14 adequately represented in models and upscaling to large-scale remains a significant challenge.

### 15 16 17 5.2.3.5 N<sub>2</sub>O budget

18  
19  
20 **[START FIGURE 5.17 HERE]**

21  
22 **Figure 5.17: Global nitrous oxide (N<sub>2</sub>O) budget (2007–2016).** Values and data sources as in Table 5.3. The  
23 atmospheric stock is calculated from mean N<sub>2</sub>O concentration, multiplying a factor of  $4.79 \pm 0.05$  Tg ppb<sup>-1</sup>  
24 (Prather et al., 2012). Pool sizes for the other reservoirs are largely unknown. Further details on data  
25 sources and processing are available in the chapter data table (Table 5.SM.6).

26  
27 **[END FIGURE 5.17 HERE]**

28  
29  
30 The synthesis of bottom-up estimates of N<sub>2</sub>O sources (Sections 5.2.3.2–5.2.3.4; Figure 5.17) yields a global  
31 source of 17.0 (12.2–23.5) TgN yr<sup>-1</sup> for the years 2007–2016 (Table 5.3). This estimate is comparable to  
32 AR5, but the uncertainty range has been reduced primarily due to improved estimates of ocean and  
33 anthropogenic N<sub>2</sub>O sources. Since AR5 (WGI, 6.4.3), improved capacity to estimate N<sub>2</sub>O sources from  
34 atmospheric N<sub>2</sub>O measurements by inverting models of atmospheric transport provides a new and  
35 independent constraint for the global N<sub>2</sub>O budget (Saikawa et al., 2014; Thompson et al., 2019; Tian et al.,  
36 2020). The decadal mean source derived from these inversions is remarkably consistent with the bottom-up  
37 global N<sub>2</sub>O budget for the same period, however, the split between land and ocean sources based on  
38 atmospheric inversions is less well constrained, yielding a smaller land source of 11.3 (10.2–13.2) TgN yr<sup>-1</sup>  
39 and a larger ocean source of 5.7 (3.4–7.2) TgN yr<sup>-1</sup>, respectively, compared to bottom-up estimates.

40  
41 Supported by multiple studies and extensive observational evidence (Sections 5.2.3.2–5.2.3.4; Figure 5.17),  
42 anthropogenic emissions contributed about 40% (7.3; uncertainty range: 4.2–11.4 TgN yr<sup>-1</sup>) to the total N<sub>2</sub>O  
43 source in 2007–2016 (*high confidence*). This estimate is larger than in AR5 (WGI, 6.4.3) due to a larger  
44 estimated effect of nitrogen deposition on soil N<sub>2</sub>O emission and the explicit consideration of the role of  
45 anthropogenic nitrogen in determining inland water and estuary emissions.

46  
47 Based on bottom-up estimates, anthropogenic emissions from agricultural nitrogen use, industry and other  
48 indirect effects have increased by 1.7 (1.0–2.7) TgN yr<sup>-1</sup> between the decades 1980–1989 and 2007–2016,  
49 and are the primary cause of the increase in the total N<sub>2</sub>O source (*high confidence*). Atmospheric inversions  
50 indicate that changes in surface emissions rather than in the atmospheric transport or sink of N<sub>2</sub>O are the  
51 cause for the increased atmospheric growth rate of N<sub>2</sub>O (*robust evidence, high agreement*) (Thompson et al.,  
52 2019). However, the increase of 1.6 (1.4–1.7) TgN yr<sup>-1</sup> in global emissions between 2000–2005 and 2010–  
53 2015 based on atmospheric inversions is somewhat larger than bottom-up estimates over the same period,  
54 primarily because of differences in the estimates of land-based emissions.

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[START TABLE 5.3 HERE]

**Table 5.3: Global N<sub>2</sub>O budget (units TgN yr<sup>-1</sup>) averaged over the 1980s, 1990s, 2000s as well as the recent decade starting in 2007.** Uncertainties represent the assessed range of source/sink estimates. All numbers are reproduced from (Tian et al., 2020) based on a compilation of inventories, bottom-up models, as well as atmospheric inversions. For detailed information on datasets, see Data Table 5.SM.6.

	AR6 1980–1989	AR6 1990–1999	AR6 2000–2009	AR6 (2007–2016)	AR5 (2006/2011)
<b>Bottom-up budget</b>					
<b>Anthropogenic Sources</b>					
Fossil Fuel combustion and Industry	0.9 (0.8–1.1)	0.9 (0.9–1.0)	1.0 (0.8–1.0)	1.0 (0.8–1.1)	0.7 (0.2–1.8)
Agriculture (incl. Aquaculture)	2.6 (1.8–4.1)	3.0 (2.1–4.8)	3.4 (2.3–5.2)	3.8 (2.5–5.8)	4.1 (1.7–4.8)
Biomass and biofuel burning	0.7 (0.7–0.7)	0.7 (0.6–0.8)	0.6 (0.6–0.6)	0.6 (0.5–0.8)	0.7 (0.2–1.0)
Wastewater	0.2 (0.1–0.3)	0.3 (0.2–0.4)	0.3 (0.2–0.4)	0.4 (0.2–0.5)	0.2 (0.1–0.3)
Inland water, estuaries, coastal zones	0.4 (0.2–0.5)	0.4 (0.2–0.5)	0.4 (0.2–0.6)	0.5 (0.2–0.7)	
Atmospheric nitrogen deposition on ocean	0.1 (0.1–0.2)	0.1 (0.1–0.2)	0.1 (0.1–0.2)	0.1 (0.1–0.2)	0.2 (0.1–0.4)
Atmospheric nitrogen deposition on land	0.6 (0.3–1.2)	0.7 (0.4–1.4)	0.7 (0.4–1.3)	0.8 (0.4–1.4)	0.4 (0.3–0.9)
Other indirect effects from CO <sub>2</sub> , climate and land-use change	0.1 (-0.4–0.7)	0.1 (-0.5–0.7)	0.2 (-0.4–0.9)	0.2 (-0.6–1.1)	
<b>Total Anthropogenic</b>	<b>5.6 (3.6–8.7)</b>	<b>6.2 (3.9–9.6)</b>	<b>6.7 (4.1–10.3)</b>	<b>7.3 (4.2–11.4)</b>	<b>6.3 (2.6–9.2)</b>
<b>Natural Sources and Sinks</b>					
Rivers, estuaries, and coastal zones	0.3 (0.3–0.4)	0.3 (0.3–0.4)	0.3 (0.3–0.4)	0.3 (0.3–0.4)	0.6 (0.1–2.9)
Open oceans	3.6 (3.0–4.4)	3.5 (2.8–4.4)	3.5 (2.7–4.3)	3.4 (2.5–4.3)	3.8 (1.8–9.4)
Soils under natural vegetation	5.6 (4.9–6.6)	5.6 (4.9–6.5)	5.6 (5.0–6.5)	5.6 (4.9–6.5)	6.6 (3.3–9.0)
Atmospheric chemistry	0.4 (0.2–1.2)	0.4 (0.2–1.2)	0.4 (0.2–1.2)	0.4 (0.2–1.2)	0.6 (0.3–1.2)
Surface sink	-0.01 (-0.3–0)	-0.01 (-0.3–0)	-0.01 (-0.3–0)	-0.01 (-0.3–0)	-0.01 (-1–0)
<b>Total natural</b>	<b>9.9 (8.5–12.2)</b>	<b>9.8 (8.3–12.1)</b>	<b>9.8 (8.2–12.0)</b>	<b>9.7 (8.0–12.0)</b>	<b>11.6 (5.5–23.5)</b>
<b>Total bottom-up source</b>	<b>15.5 (12.1–20.9)</b>	<b>15.9 (12.2–21.7)</b>	<b>16.4 (12.3–22.4)</b>	<b>17.0 (12.2–23.5)</b>	<b>17.9 (8.1–30.7)</b>
<b>Observed growth rate</b>			<b>3.7 (3.7–3.7)</b>	<b>4.5 (4.3–4.6)</b>	<b>3.6 (3.5–3.8)</b>
<b>Inferred stratospheric sink</b>			<b>12.9 (12.2–13.5)</b>	<b>13.1 (12.4–13.6)</b>	<b>14.3 (4.3–28.7)</b>
<b>Atmospheric inversion</b>					
Atmospheric loss			12.1 (11.4–13.3)	12.4 (11.7–13.3)	
Total source			15.9 (15.1–16.9)	16.9 (15.9–17.7)	
Imbalance			3.6 (2.2–5.7)	4.2 (2.4–6.4)	

9

[END TABLE 5.3 HERE]

10

#### 5.2.4 The Relative Importance of CO<sub>2</sub>, CH<sub>4</sub>, and N<sub>2</sub>O

The total influence of anthropogenic greenhouse gases (GHGs) on the Earth's radiative balance is driven by the combined effect of those gases, and the three most important were discussed separately in the previous sections. This section compares the balance of the sources and sinks of these three gases and their regional net flux contributions to the radiative forcing. CO<sub>2</sub> has multiple residence times in the atmosphere from one year to many thousands of years (Box 6.1 in Ciais et al. (2013)), and N<sub>2</sub>O has a mean lifetime of 116 years. They are both long-lived GHGs, while CH<sub>4</sub> has a lifetime of 9.0 years and is considered a short-lived GHG (see Chapter 2 for lifetime of GHGs, Chapter 6 for CH<sub>4</sub> chemical lifetime, and Chapter 7 for effective radiative forcing of all GHGs).

Figure 5.18 shows the contribution to radiative forcing of CO<sub>2</sub>, CH<sub>4</sub>, N<sub>2</sub>O, and the halogenated species since the 1900s and the more recent decades. For the period 1960–2019, the relative contribution to the total effective radiative forcing (ERF) was 63% for CO<sub>2</sub>, 11% for CH<sub>4</sub>, 6% for N<sub>2</sub>O, and 17% for the halogenated species (Chapter 7; Figure 5.18). The systematic decline in the relative contribution to ERF for CH<sub>4</sub> since 1850 is caused by slower increase rate of CH<sub>4</sub> in the recent decades, at 6, 10 and 5 ppb yr<sup>-1</sup> during 1850–2019, 1960–2019 and 2000–2019, respectively, in comparison with the increasing rate of CO<sub>2</sub> (at 0.7, 1.6 and 2.2 ppm yr<sup>-1</sup>, respectively) and N<sub>2</sub>O (at 0.4, 0.7 and 0.9 ppb yr<sup>-1</sup>, respectively) (Figure 5.4). Owing to the shorter lifetime of CH<sub>4</sub>, the effect of reduction in emission increase rate on the ERF increase is evident at inter-decadal timescales.

[START FIGURE 5.18 HERE]

**Figure 5.18: Contributions of carbon dioxide (CO<sub>2</sub>), methane (CH<sub>4</sub>), nitrous oxide (N<sub>2</sub>O) and halogenated species to the total effective radiative forcing (ERF) increase since 1850 and 1960, and for 2000 to 2009.** ERF data are taken from Annex III (based on calculations from Chapter 7). Note that the sum of the ERFs exceeds 100% because there are negative ERFs due to aerosols and clouds. Further details on data sources and processing are available in the chapter data table (Table 5.SM.6).

[END FIGURE 5.18 HERE]

Atmospheric abundance of GHGs is proportional to their emissions-loss budgets in the Earth's environment. There are multiple metrics to evaluate the relative importance of different GHGs for the global atmospheric radiation budget and the socioeconomic impacts (Section 7.6). Metrics for weighting emissions are further developed in the AR6 of IPCC WGIII. Figure 5.19 shows the regional emissions of the three main GHGs. For East Asia, Europe, Temperate North America and West Asia, the most dominant GHG source is CO<sub>2</sub> (*high confidence*) (Figure 5.19), while for East Asia, South Asia, Southeast Asia, Tropical South America, Temperate North America and Central Africa is CH<sub>4</sub> (Figure 5.19). The N<sub>2</sub>O emissions are dominant in regions with intense use of nitrogen fertilisers in agriculture. Only boreal North America showed net sinks of CO<sub>2</sub>, while close to flux neutrality is observed for North Asia, Southern Africa, and Australasia. Persistent emission of CO<sub>2</sub> is observed for Tropical and South America, northern Africa, and southeast Asia (*medium confidence*). The *medium confidence* arises from large uncertainties in the estimated non-fossil fuel CO<sub>2</sub> fluxes over these regions due to the lack of high-quality atmospheric measurements.

[START FIGURE 5.19 HERE]

**Figure 5.19: Regional distributions of net fluxes of carbon dioxide (CO<sub>2</sub>), methane (CH<sub>4</sub>), nitrous oxide (N<sub>2</sub>O) on the Earth's surface.** The region divisions, shown as the shaded map, are made based on ecoclimatic characteristics of the land. The fluxes include those from anthropogenic activities and natural causes that result from responses to anthropogenic greenhouse gases and climate change (feedbacks) as in the three budgets shown in Sections 5.2.1.5, 5.2.2.5, and 5.2.3.5. The CH<sub>4</sub> and N<sub>2</sub>O emissions are weighted by

1 arbitrary factors of 50 and 500, respectively, for depiction by common y-axes. Fluxes are shown as the  
2 mean of the inverse models as available from (Thompson et al., 2019; Friedlingstein et al., 2020; Saunio  
3 et al., 2020). Further details on data sources and processing are available in the chapter data table (Table  
4 5.SM.6).  
5

6 **[END FIGURE 5.19 HERE]**  
7  
8

### 9 **5.3 Ocean Acidification and Deoxygenation**

10  
11 The surface ocean has absorbed a quarter of all anthropogenic CO<sub>2</sub> emissions mainly through physical-  
12 chemical processes (McKinley et al., 2016; Gruber et al., 2019b; Friedlingstein et al., 2020). Once dissolved  
13 in seawater, CO<sub>2</sub> reacts with water and forms carbonic acid. In turn carbonic acid dissociates, leading to a  
14 decrease in the concentration of carbonate (CO<sub>3</sub><sup>2-</sup>) ions, and increasing both bicarbonate (HCO<sub>3</sub><sup>-</sup>) and  
15 hydrogen (H<sup>+</sup>) ion concentration, which has caused a shift in the carbonate chemistry towards a less basic  
16 state, commonly referred to as ocean acidification (Caldeira and Wickett, 2003; Orr et al., 2005; Doney et  
17 al., 2009). Although the societal concern for this problem is relatively recent (about the last 20 years), the  
18 physical-chemical basis for the ocean absorption (sink) of atmospheric CO<sub>2</sub> has been discussed much earlier  
19 by Revelle and Suess (1957). The AR5 and SROCC assessments were of *robust evidence* that the H<sup>+</sup> ion  
20 concentration is increasing in the surface ocean, thereby reducing seawater pH (= -log[H<sup>+</sup>]) (Orr et al., 2005;  
21 Feely et al., 2009; Ciais et al., 2013; Bindoff et al., 2019; Chapter 2, Section 2.3.4.1), and there is *high*  
22 *confidence* that ocean acidification is impacting marine organisms (Bindoff et al., 2019).  
23

24 Ocean oxygen decline, or deoxygenation, is driven by changes in ocean ventilation and solubility (Bindoff et  
25 al., 2019). It is *virtually certain* that anthropogenic forcing has made a substantial contribution to the ocean  
26 heat content increase over the historical period (Bindoff et al., 2019; IPCC, 2019c) (Chapter 9, Section  
27 2.3.3.1), strengthening upper water column stratification. Ocean warming decreases the solubility of  
28 dissolved oxygen in seawater, and it contributes to about 15% of the dissolved oxygen decrease in the oceans  
29 according to estimates based on solubility and the recent SROCC assessment (*medium confidence*),  
30 especially in sub-surface waters, between 100–600 m depth (Helm et al., 2011; Schmidtko et al., 2017;  
31 Breitburg et al., 2018; Oschlies et al., 2018) (SROCC, Section 5.3.1). Stratification reduces the ventilation  
32 flux into the ocean interior, contributing to most of the remaining ocean deoxygenation (Schmidtko et al.,  
33 2017; Breitburg et al., 2018) (Section 3.6.2). Deoxygenation may enhance the emissions of nitrous oxide,  
34 especially from oxygen minimum zones (OMZs) or hypoxic coastal areas (Breitburg et al., 2018; Oschlies et  
35 al., 2018). Since SROCC (Bindoff et al., 2019), CMIP6 model simulation results agree with the reported 2%  
36 loss ( $4.8 \pm 2.1$  Pmoles O<sub>2</sub>) in total dissolved oxygen in the upper ocean layer (100–600 m) for the 1970–2010  
37 period (Helm et al., 2011; Ito et al., 2017; Schmidtko et al., 2017; Kwiatkowski et al., 2020) (Section  
38 2.3.4.2). The response of marine organisms to the coupled effects of ocean warming, acidification and  
39 deoxygenation occur at different metabolic levels on different groups, and include respiratory stress and  
40 reduction of thermal tolerance by organisms (Gruber, 2011; Bindoff et al., 2019; IPCC, 2019c; Kawahata et  
41 al., 2019). An assessment of these effects on marine biota is found in WGII AR6 Chapter 2.  
42

43 This section assesses past events of ocean acidification and deoxygenation (Section 5.3.1), the historical  
44 trends and spatial variability for the upper ocean (Section 5.3.2) and the ocean interior (Section 5.3.3). Future  
45 projections for ocean acidification and the drivers in the coastal ocean are assessed in Sections 5.3.4 and  
46 5.3.5, respectively.  
47

#### 48 **5.3.1 Paleoclimate Context**

##### 49 **5.3.1.1 Paleocene-Eocene Thermal Maximum**

50  
51 The Paleocene-Eocene thermal maximum (PETM) was an episode of global warming exceeding pre-  
52 industrial temperatures by 4°C–8°C (McInerney and Wing, 2011; Dunkley Jones et al., 2013) that occurred  
53 55.9–55.7 Ma. The PETM involved a large pulse of geologic CO<sub>2</sub> released into the ocean-atmosphere  
54  
55

1 system in 3–20 kyr (Zeebe et al., 2016; Gutjahr et al., 2017; Kirtland Turner et al., 2017; Kirtland Turner,  
2 2018; Gingerich, 2019) (5.2.1.1). In response to carbon emissions during the PETM, observationally-  
3 constrained model simulations report an increase in atmospheric CO<sub>2</sub> concentrations ranging from about 900  
4 ppm to >2000 ppm (Gutjahr et al., 2017; Cui & Schubert, 2018; Anagnostou et al., 2020) (Chapter 2). The  
5 PETM thus provides a test for our understanding of the ocean’s response to the increase in carbon (and heat)  
6 emissions over geologically short timescales.  
7

8 A limited number of independent proxy records indicate that the PETM was associated with a surface ocean  
9 pH decline ranging from 0.15 to 0.30 units (Gutjahr et al., 2017; Penman et al., 2014; Babila et al., 2018). It  
10 was also accompanied by a rapid (<10 ka) shallowing of the carbonate saturation horizon, resulting in the  
11 widespread dissolution of sedimentary carbonate, followed by a gradual (100 kyr) recovery (Zachos et al.,  
12 2005; Bralower et al., 2018). The remarkable similarity among sedimentary records spanning a wide range of  
13 ecosystems suggests with *medium confidence* that the perturbation in the ocean carbonate saturation was  
14 global (Babila et al., 2018) and directly resulted from elevated atmospheric CO<sub>2</sub> levels. The degree of  
15 acidification is similar to the 0.4 pH unit decrease projected for the end of the 21st century under RCP8.5  
16 (Gattuso et al., 2015) and is estimated to have occurred at a rate about one order of magnitude slower than  
17 the current rate of ocean acidification (Zeebe et al., 2016). There is *low confidence* in the inferred rates of  
18 ocean acidification inherent to the range of uncertainties affecting rates estimates based on marine sediments  
19 (Section 5.1.2.1).  
20

21 Recent model outputs as well as globally distributed geochemical data reveal with *medium confidence*  
22 widespread ocean deoxygenation during the PETM (Dickson et al., 2012; Winguth et al., 2012; Dickson et  
23 al., 2014; Chang et al., 2018; R Emmelzwaal et al., 2019), with parts of the ocean potentially becoming  
24 drastically oxygen-depleted (anoxic) (Yao et al., 2018; Clarkson et al., 2021). Deoxygenation affected the  
25 surface ocean globally (including the Arctic Ocean) (Sluijs et al., 2006), due to vertical and lateral expansion  
26 of Oxygen Minimum Zones (OMZs) (Zhou et al., 2014) that resulted from warming and related changes in  
27 ocean stratification. Expansion of OMZs may have stimulated N<sub>2</sub>O production through water-column  
28 (de)nitrification (Junium et al., 2018). The degree to which N<sub>2</sub>O production impacted PETM warming,  
29 however, has not yet been established.  
30

31 The feedbacks associated with recovery from the PETM are uncertain, yet could include drawdown  
32 associated with silicate weathering (Zachos et al., 2005) and regrowth of terrestrial and marine organic  
33 carbon stocks (Bowen and Zachos, 2010; Gutjahr et al., 2017).  
34  
35

### 36 5.3.1.2 Last Deglacial Transition

37  
38 The Last deglacial transition (LDT) is the best documented climatic transition in the past associated with a  
39 substantial atmospheric CO<sub>2</sub> rise ranging from 190 to 265 ppm between 18–11 ka (Marcott et al., 2014). The  
40 amplitude of the deglacial CO<sub>2</sub> rise is thus on the order of magnitude of the increase undergone since the  
41 industrial revolution.  
42

43 Boron isotope ( $\delta^{11}\text{B}$ ) data suggest a 0.15–0.05 unit decrease in sea-surface pH (Hönisch and Hemming,  
44 2005; Henehan et al., 2013) across the LDT, an average rate of decline of about 0.002 units per century  
45 compared with the current rate of more than 0.1 units per century (Bopp et al., 2013; Gattuso et al., 2015).  
46 Planktonic foraminiferal shell weights decreased by 40% to 50% (Barker & Elderfield, 2002), and coccolith  
47 mass decreased by about 25% (Beaufort et al., 2011) across the LDT. Independent proxy reconstructions  
48 thus highlight with *high confidence* that pH values decreased as atmospheric CO<sub>2</sub> concentrations increased  
49 across the LDT. There is however *low confidence* in the inferred rate of ocean acidification owing to  
50 multiple sources of uncertainties affecting rates estimates based on marine sediments (Section 5.1.2.1).  
51 Geochemical and micropaleontological evidence suggest that intermediate-depth OMZs almost vanished  
52 during the LGM (Jaccard et al., 2014). However, multiple lines of evidence suggest with *medium confidence*  
53 that the deep (>1500 m) ocean became depleted in O<sub>2</sub> (concentrations were possibly lower than 50  $\mu\text{mol kg}^{-1}$ )  
54 globally (Jaccard and Galbraith, 2012; Hoogakker et al., 2015, 2018, Gottschalk et al., 2016, 2020;  
55 Anderson et al., 2019) as a combined result of sluggish ventilation of the ocean subsurface (Gottschalk et al.,



2016, 2020; Skinner et al., 2017) and a generally more efficient marine biological carbon pump (Buchanan et al., 2016; Yamamoto et al., 2019; Galbraith and Skinner, 2020).

During the LDT, deep ocean ventilation increased as Antarctic bottom water (AABW) (Skinner et al., 2010; Gottschalk et al., 2016; Jaccard et al., 2016) and subsequently the Atlantic meridional overturning circulation (McManus et al., 2004; Lippold et al., 2016) resumed, transferring previously sequestered remineralised carbon from the ocean interior to the upper ocean and eventually the atmosphere (Skinner et al., 2010; Galbraith and Jaccard, 2015; Gottschalk et al., 2016; Ronge et al., 2016; Sikes et al., 2016; Rae et al., 2018; Ronge et al., 2020), contributing to the deglacial CO<sub>2</sub> rise. Intermediate depths lost oxygen as a result of sluggish ventilation and increasing temperatures (decreasing saturation) as the world emerged from the last Glacial period OMZs underwent a large volumetric increase at the beginning of the Bølling-Allerød (B/A), a northern-hemisphere wide warming event, 14.7 ka (Jaccard and Galbraith, 2012; Praetorius et al., 2015) with deleterious consequences for benthic ecosystems (e.g. Moffitt et al., 2015). These observations indicate with *high confidence* that the rate of warming, affecting the solubility of oxygen and upper water column stratification, coupled with changes in subsurface ocean ventilation impose a direct control on the degree of ocean deoxygenation, implying a high sensitivity of ocean oxygen loss to warming. The expansion of OMZs contributed to a widespread increase in water column (de)nitrification (Galbraith et al., 2013), which contributed substantially to enhanced marine N<sub>2</sub>O emissions (Schilt et al., 2014; Fischer et al., 2019). Nitrogen stable isotope measurements on N<sub>2</sub>O extracted from ice cores suggest that approximately one third (on the order of  $0.7 \pm 0.3 \text{ TgN yr}^{-1}$ ) of the deglacial increase in N<sub>2</sub>O emissions relates to oceanic sources (Schilt et al., 2014; Fischer et al., 2019).

### 5.3.2 Historical Trends and Spatial Characteristics in the Upper Ocean

#### 5.3.2.1 Reconstructed Centennial Ocean Acidification Trends

Ocean pH timeseries are based on the reconstruction of coral boron isotope ratios ( $\delta^{11}\text{B}$ ). A majority of coral  $\delta^{11}\text{B}$  data have been generated from the western Pacific region with a few records from the Atlantic Ocean. Biweekly resolution paleo-pH records show monsoonal variation of about 0.5 pH unit in the South China Sea (Liu et al., 2014). Interannual ocean pH variability in the range of 0.07–0.16 pH unit characterises southwest Pacific corals that are attributed to ENSO (Wu et al., 2018a) and river runoff (D’Olivo et al., 2015). Decadal (10, 22 and 48-year) ocean pH variations in the southwest Pacific have been linked to the Interdecadal Pacific Oscillation, causing variations of up to 0.30 pH unit in the Great Barrier Reef (Pelejero et al., 2005; Wei et al., 2009) but weaker (about 0.08 pH unit) in the open ocean (Wu et al., 2018a). Decadal variations in the South China Sea ocean pH changes of 0.10–0.20 also have been associated with the variation in the East Asian monsoon (Liu et al., 2014; Wei et al., 2015), as a weakening of the Asian winter monsoon leads to sluggish water circulation within the reefs, building up localised CO<sub>2</sub> concentration in the water due to calcification and respiration.

Since the beginning of the industrial period in the mid-19th century, coral  $\delta^{11}\text{B}$ -derived ocean pH has decreased by 0.06–0.24 pH unit in the South China Sea (Liu et al., 2014; Wei et al., 2015) and 0.12 pH unit in the southwest Pacific (Wu et al., 2018a). Since the mid-20th century, a distinct feature of coral  $\delta^{11}\text{B}$  records relates to ocean acidification trends, albeit having a wide-range of values: 0.12–0.40 pH unit in the great barrier reef (Wei et al., 2009; D’Olivo et al., 2015), 0.05–0.08 pH unit in the northwest Pacific (Shinjo et al., 2013) and 0.04–0.09 pH unit in the Atlantic Ocean (Goodkin et al., 2015; Fowell et al., 2018). Concurrent coral carbon isotopic ( $\delta^{13}\text{C}$ ) measurements infer ocean uptake of anthropogenic CO<sub>2</sub> from the combustion of fossil fuel, based on the lower abundance of <sup>13</sup>C in fossil fuel carbon. Western Pacific coral records show depleted  $\delta^{13}\text{C}$  trends since the late 19th century that are more prominent since the mid-20th century (*high confidence*) (Pelejero, 2005; Wei et al., 2009; Shinjo et al., 2013; Liu et al., 2014; Kubota et al., 2017; Wu et al., 2018).

Overall, many of the records show a highly variable seawater pH underlying strong imprints of internal climate variability (*high confidence*), and in most instances superimposed on a decreasing  $\delta^{11}\text{B}$  trend that is indicative of anthropogenic ocean acidification in recent decades (*medium confidence*). The robustness of

1 seawater pH reconstructions is currently limited by the uncertainty on the calibration of the  $\delta^{11}\text{B}$  proxy in  
2 different tropical coral species.

### 5.3.2.2 Observations of Ocean Acidification over the Recent Decades

7 SROCC (Section 5.2.2.3) indicated it is *virtually certain* that the ocean has undergone acidification globally  
8 in response to ocean  $\text{CO}_2$  uptake and concluded that pH in open ocean surface water has changed by a  
9 *virtually certain* range of  $-0.017$  to  $-0.027$  pH units per decade since the late 1980s. Since SROCC,  
10 continued observations of seawater carbonate chemistry at ocean time series stations and compiled shipboard  
11 studies providing temporally resolved and methodologically consistent datasets have further strengthened the  
12 evidence of the progress of acidification across all regions of the oceans (Jiang et al., 2019) (Figure 5.20;  
13 Supplementary Material Table 5.SM.3) (Section 2.3.3.5).

15 In the subtropical open oceans, decreases in pH have been reported with a *very likely* range of rate from  $-$   
16  $0.016$  to  $-0.019$  pH units per decade since 1980s, which equate to approximately 4 % increase in hydrogen  
17 ion concentration ( $[\text{H}^+]$ ) per decade. Accordingly, the saturation state  $\Omega$  ( $=[\text{Ca}^{2+}][\text{CO}_3^{2-}]/K_{\text{sp}}$ ) of seawater  
18 with respect to calcium carbonate mineral aragonite has been declining at rates ranging from  $-0.07$  to  $-0.12$   
19 per decade (González-Dávila et al., 2010; Feely et al., 2012; Bates et al., 2014; Takahashi et al., 2014; Ono  
20 et al., 2019; Bates and Johnson, 2020) (Supplementary Material Table 5.SM.3). These rates are consistent  
21 with the rates expected from the transient equilibration with increasing atmospheric  $\text{CO}_2$  concentrations, but  
22 the variability of rate in decadal time-scale has also been detected with *robust evidence* (Ono et al., 2019;  
23 Bates and Johnson, 2020). In the tropical Pacific, its central and eastern upwelling zones exhibited a faster  
24 pH decline of  $-0.022$  to  $-0.026$  pH unit per decade due to increased upwelling of  $\text{CO}_2$ -rich sub-surface  
25 waters in addition to anthropogenic  $\text{CO}_2$  uptake (Sutton et al., 2014; Lauvset et al., 2015). By contrast, warm  
26 pool in the western tropical Pacific exhibited slower pH decline of  $-0.010$  to  $-0.013$  pH unit per decade  
27 (Supplementary Material Table 5.SM.3) (Lauvset et al., 2015; Ishii et al., 2020). Observational and modeling  
28 studies (Nakano et al., 2015; Ishii et al., 2020) consistently suggest that slower acidification in this region is  
29 attributable to the anthropogenic  $\text{CO}_2$  taken up in the extra-tropics around a decade ago and transported to  
30 the tropics via shallow meridional overturning circulations.

32 In open subpolar and polar zones, the *very likely* range ( $-0.003$  to  $-0.026$  pH unit per decade) and  
33 uncertainty (up to 0.010) observed in pH decline are larger than in the subtropics, reflecting the complex  
34 interplay between physical and biological forcing mechanisms (Olafsson et al., 2009; Midorikawa et al.,  
35 2012; Bates et al., 2014; Takahashi et al., 2014; Lauvset et al., 2015; Wakita et al., 2017). Nevertheless, the  
36 *high agreement* of pH decline among these available time-series studies leads to *high confidence* in the trend  
37 of acidification in these zones. In the Arctic Ocean, a temporally limited time series of carbonate chemistry  
38 measurements prevents drawing robust conclusions on ocean acidification trends. However, the carbonate  
39 saturation state  $\Omega$  is generally low, and observational studies show with *robust evidence* that the recent  
40 extensive melting of sea ice leading to enhanced air-sea  $\text{CO}_2$  exchange, large freshwater inputs, together with  
41 river discharge and glacial drainage, as well as the degradation of terrestrial organic matter in seawater,  
42 result in the decline of  $\Omega$  of aragonite to undersaturation (Bates et al., 2009; Chierici and Fransson, 2009;  
43 Yamamoto-Kawai et al., 2009; Azetsu-Scott et al., 2010; Robbins et al., 2013; Fransson et al., 2015;  
44 Semiletov et al., 2016; Anderson et al., 2017; Qi et al., 2017; Zhang et al., 2020; Beupré-Laperrière et al.,  
45 2020) (SROCC Section 3.2.1.2.4, IPCC, (2019b)). The low saturation state of aragonite ( $\Omega \sim 1$ ) has also been  
46 observed in surface waters of the Antarctic coastal zone associated with freshwater input from glacier  
47 (Mattsdotter Björk et al., 2014) and with upwelling of deep water (Hauri et al., 2015) as well as along eastern  
48 boundary upwelling systems (Feely et al., 2016a).

50 Overall, in agreement with SROCC, it is *virtually certain* from these observational studies that ocean surface  
51 waters undergo acidification globally with the  $\text{CO}_2$  increase in the atmosphere. These sustained  
52 measurements over the past decades and campaign studies of ocean carbonate chemistry also highlight with  
53 *robust evidence* that trends of acidification have been modulated by the variability and changes in physical  
54 and chemical states of ocean including those affected by the warming of the cryosphere, requiring their  
55 improved understandings.

1  
2  
3 [START FIGURE 5.20 HERE]  
4

5 **Figure 5.20: Multi-decadal trends of pH (Total Scale) in surface layer at various sites of the oceans and a global**  
6 **distribution of annual mean pH adjusted to the year 2000.** Time-series data of pH are from Dore et  
7 al., 2009; Olafsson et al., 2009; González-Dávila et al., 2010; Bates et al., 2014b; Takahashi et al., 2014;  
8 Wakita et al., 2017; Merlivat et al., 2018; Ono et al., 2019; and Bates and Johnson, 2020. Global  
9 distribution of annual mean pH have been evaluated from data of surface ocean  $p\text{CO}_2$  measurements  
10 (Bakker et al., 2016; Jiang et al., 2019). Acronyms in panels: KNOT and K2 - Western Pacific subarctic  
11 gyre time-series; HOT - Hawaii Ocean Time-series; BATS - Bermuda Atlantic Time-series Study;  
12 DYFAMED - Dynamics of Atmospheric Fluxes in the Mediterranean Sea; ESTOC - European Station for  
13 Time-series in the Ocean Canary Islands; CARIACO - Carbon Retention in a Colored Ocean Time-series.  
14 Further details on data sources and processing are available in the chapter data table (Table 5.SM.6).  
15

16 [END FIGURE 5.20 HERE]  
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18

### 19 5.3.3 Ocean Interior Change

#### 20 5.3.3.1 Ocean Memory – Acidification in the Ocean Interior

21 Advances in observations and modelling for ocean physics and biogeochemistry and established knowledge  
22 of ocean carbonate chemistry show with *very high confidence* that anthropogenic  $\text{CO}_2$  taken up into the  
23 ocean surface layer is further spreading into the ocean interior through ventilation processes including  
24 vertical mixing, diffusion, subduction and meridional overturning circulations (Sallée et al., 2012; Bopp et  
25 al., 2015; Nakano et al., 2015; Iudicone et al., 2016; Toyama et al., 2017; Perez et al., 2018; Gruber et al.,  
26 2019b) (Sections 2.3.3.5, 5.2.1.3 and 9.2.2.3) and is causing acidification in the ocean interior. The net  
27 change in oxygen consumptions by aerobic respiration of marine organisms further influences acidification  
28 by releasing  $\text{CO}_2$  (Chen et al., 2017; Breitburg et al., 2018; Robinson, 2019) (Section 5.3.3.2).  
29

30 Basin-wide and global syntheses of ocean interior carbon observations for the past decades show that the  
31 extent of acidification due to anthropogenic  $\text{CO}_2$  invasion tends to diminish with depth (*very high*  
32 *confidence*) (Woosely et al., 2016; Lauvset et al., 2020) (Carter et al., 2017a)(Figure 5.21; Section  
33 5.2.1.3.3). The regions of deep convection such as subpolar North Atlantic and Southern Ocean present the  
34 deepest acidification detections below 2000 m (*medium confidence*). Mid-latitude zones within the  
35 subtropical cells and tropical regions present a relatively deep and shallow detection, respectively. A pH  
36 decrease has also been observed on the Antarctic continental shelf (Hauck et al., 2010; Williams et al.,  
37 2015). Acidification is also underway in the subsurface to intermediate layers of the Arctic Ocean due to the  
38 inflow of ventilated waters from the North Atlantic and the North Pacific (Qi et al., 2017; Ulfso et al.,  
39 2018).  
40  
41  
42  
43

44 [START FIGURE 5.21 HERE]  
45

46 **Figure 5.21: Spread of ocean acidification from the surface into the interior of ocean since pre-industrial times.**  
47 (a) map showing the three transects used to create the cross sections shown in (b) and (c); vertical  
48 sections of the changes in (b) pH and (c) saturation state of aragonite ( $\Omega_{\text{arag}}$ ) between 1800–2002 due to  
49 anthropogenic  $\text{CO}_2$  invasion (colour). Contour lines are their contemporary values in 2002. The red  
50 transect begins in the Nordic Seas and then follows the GO-SHIP lines A16 southward in the Atlantic  
51 Ocean, SR04 and S04P westward in the Southern Ocean, and P16 northward in the Pacific Ocean. The  
52 purple line follows the GO-SHIP line I09 southward in the Indian Ocean. The green line on the smaller  
53 inset crosses the Arctic Ocean from the Bering Strait to North Pole along  $175^\circ\text{W}$  and from the North Pole  
54 to the Fram Strait along  $5^\circ\text{E}$  (Lauvset et al., 2020). Further details on data sources and processing are  
55 available in the chapter data table (Table 5.SM.6).  
56

57 [END FIGURE 5.21 HERE]

1  
2  
3 A significant increase in acidification resulting from net metabolic CO<sub>2</sub> release coupled with ocean  
4 circulation changes has been shown with *high confidence* in large swathes of intermediate waters both in the  
5 Pacific and Atlantic oceans (Byrne et al., 2010; Carter et al., 2017; Chu et al., 2016; Dore et al., 2009; Rios  
6 et al., 2015; Lauvset et al., 2020). For example, ocean circulation contributes a pH change of  $-0.013 \pm 0.013$   
7 to the overall observed change of  $-0.029 \pm 0.014$  for 1993–2013 at depths around 1000 m at 30°S–40°S in  
8 the South Atlantic ocean (Ríos et al., 2015). Long-term repeated observations in the North Pacific show a  
9 decline in dissolved oxygen ( $-4.0 \mu\text{mol kg}^{-1}$  per decade at maximum) being sustained in the intermediate  
10 water since the 1980s (Takatani et al., 2012; Sasano et al., 2015), and thus the amplification of acidification  
11 associated with the weakening ventilation is thought to have been occurring persistently. In contrast, for the  
12 North Pacific subtropical mode water, large decadal variability in pH and aragonite saturation state with  
13 amplitudes of about 0.02 and about 0.1, respectively, are superimposed on secular declining trends due to  
14 anthropogenic CO<sub>2</sub> invasion (Oka et al., 2019). This is associated with the variability in ventilation due to  
15 the approximately 50% variation in the formation volume of the mode water that is forced remotely by the  
16 Pacific decadal oscillation (Qiu et al., 2014; Oka et al., 2015).

17  
18 These trends of acidification in the ocean interior leads to *high confidence* in shoaling of the saturation  
19 horizons of calcium carbonate minerals where  $\Omega = 1$ . In the Pacific Ocean where the aragonite saturation  
20 horizon is shallower (a few hundred meters to 1200 m; Figure 5.21c), the rate of its shoaling is on the order  
21 of 1–2 m yr<sup>-1</sup> (Feely et al., 2012; Ross et al., 2020). In contrast, shoaling rates of 4 m yr<sup>-1</sup> to 1710 m for  
22 1984–2008 and of 10–15 m yr<sup>-1</sup> to 2250 m for 1991–2016 have been observed in the Iceland sea and the  
23 Irminger sea, respectively (Olafsson et al., 2009; Perez et al., 2018).

24  
25 In summary, ocean acidification is spreading into the ocean interior. Its rates at depths are controlled by the  
26 ventilation of the ocean interior as well as anthropogenic CO<sub>2</sub> uptake at the surface, thereby diminishing with  
27 depth (*very high confidence*) (Figure 5.21). Variability in ocean circulation modulates the trend of ocean  
28 acidification at depths through the changes in ventilation and their impacts on metabolic CO<sub>2</sub> content, but  
29 there are large knowledge gaps of ventilation changes leading to *low confidence* in their impacts in many  
30 ocean regions (Sections 9.2.2.3 and 9.3.2; Section 5.3.3.2).

### 31 32 33 5.3.3.2 Ocean Deoxygenation and its Implications for GHGs

34  
35 As summarised in SROCC (Section 5.2.2.4), there is a growing consensus that between 1970–2010 the open  
36 ocean has *very likely* lost 0.5–3.3% of its dissolved oxygen in the upper 1000 m depth (Helm et al., 2011; Ito  
37 et al., 2017; Schmidtko et al., 2017; Bindoff et al., 2019) (Section 2.3.3.6). Regionally, the equatorial and  
38 North Pacific, the Southern Ocean and the South Atlantic have shown the greatest oxygen loss of up to 30  
39 mol m<sup>-2</sup> per decade (Schmidtko et al., 2017). Warming – via solubility reduction and circulation changes –,  
40 mixing and respiration are considered the major drivers with 50% of the oxygen loss for the upper 1000 m of  
41 the global oceans attributable to the solubility reduction (Schmidtko et al., 2017). Climate variability also  
42 modifies the oxygen loss on interannual and decadal timescales especially for the tropical ocean OMZs  
43 (Deutsch et al., 2011, 2014; Llanillo et al., 2013) and the North Pacific subarctic zone (Whitney et al., 2007;  
44 Sasano et al., 2018; Cummins and Ross, 2020). However, quantifying the oxygen decline and variability and  
45 attributing them to processes in different regions remains challenging (Oschlies et al., 2018; Levin, 2018).  
46 ESMs in coupled model intercomparison project phase 5 (CMIP5) and CMIP6 corroborate the decline in  
47 ocean oxygen, and project a continuing and accelerating decline with a strong impact of natural climate  
48 variability under high emission scenarios (Bopp et al., 2013; Kwiatkowski et al., 2020; Long et al., 2016).  
49 However, CMIP5 models did not reproduce observed patterns for oxygen changes in the tropical thermocline  
50 and generally simulated only about half the oxygen loss inferred from observations (Oschlies et al., 2018).  
51 CMIP6 models have more realistic simulated mean state of ocean biogeochemistry than CMIP5 models due  
52 to improved ocean physical processes and better representation of biogeochemical processes (Séférian et al.,  
53 2020). They also exhibit enhanced ocean warming as a result of an increase in the equilibrium climate  
54 sensitivity (ECS) of CMIP6 relative to CMIP5 models, which contributes to increased stratification and  
55 reduced subsurface ventilation (4.3.1, 4.3.4, 5.3.3.2, 7.4.2, 7.5.6, 9.2.1, TS2.4). Consequently, CMIP6 model

ensembles not only reproduce the ocean deoxygenation trend of  $-0.30$  to  $-1.52$   $\text{mmol m}^{-3}$  per decade between 1970–2010 reported in SROCC (Section 5.2.2.4) with a *very likely* range, but also project 32–71 % greater subsurface (100–600 m) oxygen decline relative to their RCP analogues in CMIP5, reaching to the *likely* range of decline of  $6.4 \pm 2.9$   $\text{mmol m}^{-3}$  under SSP1–2.6 and  $13.3 \pm 5.3$   $\text{mmol m}^{-3}$  under SSP5–8.5, from 1870–1899 to 2080–2099. However, they also exhibit enhanced surface ocean warming as a result of an increased climate sensitivity (ECS), which contributes to greater reduction in subsurface ventilation. Consequently, CMIP6 model ensembles now reproduce the recent observed historical ocean deoxygenation trend of  $-0.30$  to  $-1.52$   $\text{mmol m}^{-3}$  per decade between 1970–2010 reported in SROCC (Section 5.2.2.4) within 90% confidence range, but project 32–71 % greater subsurface (100–600m) oxygen decline relative to their RCP analogues in CMIP5, reaching to the *likely* range of decline of  $6.4 \pm 2.9$   $\text{mmol m}^{-3}$  under SSP1–2.6 and  $13.3 \pm 5.3$   $\text{mmol m}^{-3}$  under SSP5–8.5, from 1870–1899 to 2080–2099 due to increased warming (Kwiatkowski et al., 2020). It is concluded that the oxygen content of subsurface ocean is projected to transition to historically unprecedented condition with decline over the 21st century (*medium confidence*).

In oxygen-depleted waters, microbial processes (denitrification and anammox, i.e. anaerobic ammonium oxidation) (Kuypers et al., 2005; Codispoti, 2007; Gruber and Galloway, 2008) remove fixed nitrogen, and thus when upwelled waters reach the photic zone, primary production becomes nitrogen-limited (Tyrrell and Lucas, 2002). However, in other oceanic regions, increased water-column stratification due to warming may reduce the amount of  $\text{N}_2\text{O}$  reaching the surface and thereby decrease  $\text{N}_2\text{O}$  flux to the atmosphere. Landolfi et al. (2017) suggest that by 2100, under the RCP8.5 scenario, total  $\text{N}_2\text{O}$  production in the ocean may decline by 5% and  $\text{N}_2\text{O}$  emissions be reduced by 24% relative to the pre-industrial era due to decreased organic matter export and anthropogenic driven changes in ocean circulation and atmospheric  $\text{N}_2\text{O}$  concentrations. Projected oxygen loss in the ocean is thought to result in an ocean-climate feedback through changes in the natural emission of greenhouse gases (*low confidence*).

The areas with relatively rapid oxygen decrease include OMZs in the tropical oceans, where oxygen content has been decreasing at a rate of  $0.9$  to  $3.4$   $\mu\text{mol kg}^{-1}$  per decade in the thermocline for the past five decades (Stramma et al., 2008). Low oxygen, low pH and shallow aragonite saturation horizons in the OMZs of the eastern boundary upwelling regions co-occur, affecting ecosystem structure (Chavez et al., 2008) and function in the water column, including the presently unbalanced nitrogen cycle (Paulmier and Ruiz-Pino, 2009). The coupling between upwelling, productivity, and oxygen depletion feeds back to biological productivity and the role of these regions as sinks or sources of climate active gases. When OMZ waters upwell and impinge on the euphotic zone, they release significant quantities of greenhouse gases, including  $\text{N}_2\text{O}$  ( $0.81$ – $1.35$   $\text{TgN yr}^{-1}$ ),  $\text{CH}_4$  ( $0.27$ – $0.38$   $\text{TgCH}_4 \text{ yr}^{-1}$ ), and  $\text{CO}_2$  (yet to be quantified) to the atmosphere, exacerbating global warming (Paulmier et al., 2008; Naqvi et al., 2010; Kock et al., 2012; Arévalo-Martínez et al., 2015; Babbín et al., 2015a; Farías et al., 2015). Modelling projections suggest a global decrease of 4 to 12% in oceanic  $\text{N}_2\text{O}$  emissions (from  $3.71$ – $4.03$   $\text{TgN yr}^{-1}$  to  $3.54$ – $3.56$   $\text{TgN yr}^{-1}$ ) from 2005 to 2100 under RCP8.5, despite a tendency to increased  $\text{N}_2\text{O}$  production in the OMZs, associated primarily with denitrification (Martínez-Rey et al., 2015). It is difficult to single out the contribution of nitrification and denitrification which can occur simultaneously. A rigorous separation of these two processes would require more mechanistic parameterisations that have been hindered by the still large conceptual and parametric uncertainties (Babbín et al., 2015; Trimmer et al., 2016; Landolfi et al., 2017). Furthermore, the correlation between  $\text{N}_2\text{O}$  and oxygen varies with microorganisms present, nutrient concentrations, and other environmental variables (Voss et al., 2013).

In summary, total oceanic  $\text{N}_2\text{O}$  emissions were projected to decline by 4–12% from 2005–2100 (Martínez-Rey et al., 2015) and by 24% from the pre-industrial era to 2100 (Landolfi et al., 2017) under RCP8.5. However, there is *low confidence* in the reduction in  $\text{N}_2\text{O}$  emission to the atmosphere, because of large conceptual and parametric uncertainties, a limited number of modelling studies that explored this process, and greater oxygen losses simulated in CMIP6 models than in CMIP5 models (Kwiatkowski et al., 2020).

### 5.3.4 Future Projections for Ocean Acidification

#### 5.3.4.1 Future Projections with Earth System Models

1  
2 Projections with CMIP5 ESMs, reported in AR5 (Section 6.4.4) and SROCC (Section 5.2.2.3; (IPCC,  
3 2019b)), showed changes in global mean surface ocean pH from 1870–1899 to 2080–2099 of  $-0.14 \pm 0.001$   
4 (inter-model standard deviation) under RCP2.6 and  $-0.38 \pm 0.005$  under RCP8.5 with pronounced regional  
5 variability (Bopp et al., 2013; Hurd et al., 2018). They also projected faster pH declines in mode waters  
6 below seasonal mixed layers (Resplandy et al., 2013; Watanabe and Kawamiya, 2017) as has been observed  
7 in the Atlantic (Salt et al., 2015) and in the Pacific (Carter et al., 2019), because of the net CO<sub>2</sub> release by  
8 respiration and lowering CO<sub>2</sub> buffering capacity of seawater. In these CO<sub>2</sub> concentration-driven simulations,  
9 the level of acidification in the surface ocean is primarily determined by atmospheric CO<sub>2</sub> concentration and  
10 regional seawater carbonate chemistry, thereby providing consistent projections across models. New  
11 projections with CMIP6 ESMs show greater surface pH decline of  $-0.16 \pm 0.002$  under the SSP1–2.6 and  
12  $-0.44 \pm 0.005$  under SSP5–8.5 from 1870–1899 to 2080–2099 (Kwiatkowski et al., 2020) (Section 4.3.2.5;  
13 Cross-Chapter Box 5.3). The greater pH declines in CMIP6 are primarily a consequence of higher  
14 atmospheric CO<sub>2</sub> concentrations in SSPs than their CMIP5-RCP analogues (Kwiatkowski et al., 2020).  
15 Ocean acidification is also projected to occur with *high confidence* in the abyssal bottom waters in regions  
16 such as the northern North Atlantic and the Southern Ocean (Sulpis et al., 2019), with the rates of global  
17 mean pH decline of  $-0.018 \pm 0.001$  under SSP1–2.6 and  $-0.030 \pm 0.002$  under SSP5–8.5 from 1870–1899 to  
18 2080–2099 in CMIP6 (Kwiatkowski et al., 2020).

19  
20 In surface ocean, changes in the amplitude of seasonal variations in pH are also projected to occur with *high*  
21 *confidence*. ESMs in CMIP6 s show  $+73 \pm 12\%$  increase in the amplitude of seasonal variation in hydrogen  
22 ion concentration ( $[H^+]$ ) but  $10 \pm 5\%$  decrease in the seasonal variation in pH ( $= -\log [H^+]$ ) from 1995–2014  
23 to 2080–2099 under SSP5–8.5. The simultaneous amplification of  $[H^+]$  and attenuation of pH seasonal  
24 cycles is counterintuitive but is the consequence of greater increase in the annual mean  $[H^+]$  due to  
25 anthropogenic CO<sub>2</sub> invasion than the corresponding increase in its seasonal amplitude. These changes are  
26 consistent with the amplification/attenuation of the seasonal variation of  $+81 \pm 16\%$  for  $[H^+]$  and  $-16 \pm 7\%$   
27 for pH from 1990–1999 to 2090–2099 under RCP8.5 in CMIP5 (Kwiatkowski and Orr, 2018).

28  
29 The signal of ocean acidification in surface ocean is large and is projected to emerge beyond the range of  
30 natural variability within the time scale of a decade in all ocean basins (Schlunegger et al., 2019). There is  
31 *high agreement* among modelling studies that the largest pH decline and large-scale undersaturation of  
32 aragonite in surface seawater start to occur first in polar oceans (Orr et al., 2005; Steinacher et al., 2009;  
33 Hurd et al., 2018; Jiang et al., 2019). Under SSP5–8.5, the largest surface pH decline, exceeding 0.45  
34 between 1995–2014 and 2080–2099, occurs in the Arctic Ocean (Kwiatkowski et al., 2020). The freshwater  
35 input from sea-ice melt is an additional factor leading to a faster decline of aragonite saturation level than  
36 expected from the anthropogenic CO<sub>2</sub> uptake (Yamamoto et al., 2012). The increase in riverine and glacial  
37 discharges that provide terrigenous carbon, nutrients and alkalinity as well as freshwater are the other factors  
38 modifying the rate of acidification in the Arctic Ocean. However, their impacts have been projected in a  
39 limited number of studies with extensive knowledge gaps and model simplifications leading to *low*  
40 *confidence* in their impacts (Terhaar et al., 2019; Hopwood et al., 2020). In the Southern Ocean, the  
41 aragonite undersaturation starts in 2030 in RCP8.5, and the area that experiences aragonite undersaturation  
42 for at least one month per year by 2100 is projected to be more than 95%. Under RCP2.6, short periods (< 1  
43 month) of aragonite undersaturation are expected to be found in less than 2% to the area during this century  
44 (Sasse et al., 2015; Hauri et al., 2016; Negrete-García et al., 2019). These long term projections are modified  
45 at interannual timescales by large-scale climate modes (Ríos et al., 2015) such as the El Niño southern  
46 oscillation and the southern annular mode (Conrad and Lovenduski, 2015). In other regions, acidification  
47 trends are influenced by a range of processes such as changes in ocean circulation, temperature, salinity,  
48 carbon cycling, and the structure of the marine ecosystem. As, at present, models do not resolve fine-scale  
49 variability of these processes, current projections do not fully capture the changes that the marine  
50 environment will experience in the future (Takeshita et al., 2015; Turi et al., 2016).

51  
52 Overall, with the rise of atmospheric CO<sub>2</sub>, the physics of CO<sub>2</sub> transfer across the air-sea interface, the  
53 carbonate chemistry in seawater, the trends of ocean acidification being observed in the past decades  
54 (Section 5.3.3.2) and modelling studies described in this section, it is *virtually certain* that ocean  
55 acidification will continue to grow. However, the magnitude and sign of many of ocean carbon-climate

1 feedbacks are still poorly constrained (Matear and Lenton, 2014; Matear and Lenton, 2018), leading to *low*  
2 *confidence* in their significant and long-lasting impacts on ocean acidification.

#### 5.3.4.2 *Reversal of Ocean Acidification by Carbon Dioxide Removal*

7 Reversing the increase in atmospheric CO<sub>2</sub> concentrations through negative emissions (Section 5.6) will  
8 reverse ocean acidification at the sea surface but will not result in rapid amelioration of ocean acidification in  
9 the deeper ocean (Section 5.3.3.2). The ocean's uptake of atmospheric CO<sub>2</sub> will start to decrease as  
10 atmospheric CO<sub>2</sub> decreases (Mathesius et al., 2015; Tokarska and Zickfeld, 2015) (Sections 5.4.5, 5.4.10;  
11 5.6.2.1) However, because of the long timescales of the ocean turnover that transfers CO<sub>2</sub> from the upper to  
12 the deep ocean, excess carbon will continue to accumulate in the deep ocean even after a decrease in  
13 atmospheric CO<sub>2</sub> (Cao et al., 2014; Mathesius et al., 2015; Tokarska and Zickfeld, 2015; Li et al., 2020).  
14 There is thus *high confidence* that CO<sub>2</sub> emissions leave a long-term legacy in ocean acidification, and are  
15 therefore irreversible at multi-human generational scales, even with aggressive atmospheric CO<sub>2</sub> removal.

#### 5.3.5 *Coastal Ocean Acidification and Deoxygenation*

20 The coastal ocean, from the shore line to the isobath of 200 m, is highly heterogeneous due to the complex  
21 interplay between physical, biogeochemical and anthropogenic factors (Gattuso et al., 1998; Chen and  
22 Borges, 2009; Dürr et al., 2011; Laruelle et al., 2014; McCormack et al., 2016). These areas, according to  
23 SROCC (Bindoff et al., 2019) are, with *high confidence*, already affected by ocean acidification and  
24 deoxygenation. This section assesses the drivers and spatial variability of acidification and deoxygenation  
25 based on new observations and data products.

##### 5.3.5.1 *Drivers*

30 Observations and data products including models (Astor et al., 2013; Bakker et al., 2016; Kosugi et al., 2016;  
31 Vargas et al., 2016; Laruelle et al., 2017, 2018; Orselli et al., 2018; Roobaert et al., 2019; Cai et al., 2020;  
32 Sun et al., 2020a) confirm the strong spatial and temporal variability in the coastal ocean surface carbonate  
33 chemistry and sea-air CO<sub>2</sub> fluxes (*high agreement, robust evidence*). The anthropogenic CO<sub>2</sub>-induced  
34 acidification is either mitigated or enhanced through biological processes; primary production removes  
35 dissolved CO<sub>2</sub> from the surface, and respiration adds CO<sub>2</sub> and consumes oxygen in the subsurface layers.  
36 The relative intensity of these processes is controlled by natural or anthropogenic eutrophication. Other  
37 drivers of variability include biological community composition, freshwater input from rivers or ice-melting,  
38 sea-ice cover and calcium carbonate precipitation/dissolution dynamics, coastal upwelling and regional  
39 circulation, and seasonal surface cooling (Fransson et al., 2015, 2017; Feely et al., 2018; Roobaert et al.,  
40 2019; Cai et al., 2020; Hauri et al., 2020; Monteiro et al., 2020b; Sun et al., 2020a). Near-shore surface  
41 waters are often supersaturated with CO<sub>2</sub>, regardless of the latitude, especially in highly populated areas  
42 receiving substantial amounts of domestic and industrial sewage (Chen and Borges, 2009). Nevertheless,  
43 thermal or haline stratified eutrophic coastal areas may act as net atmospheric CO<sub>2</sub> sinks (Chou et al., 2013;  
44 Cotovicz Jr. et al., 2015a). Continental shelves, excluding near-shore areas, act as CO<sub>2</sub> sinks at a rate of  $0.2 \pm$   
45  $0.02 \text{ PgC yr}^{-1}$  (Laruelle et al., 2014; Roobaert et al., 2019), considering ice-free areas only. Under increasing  
46 atmospheric CO<sub>2</sub> and eutrophication, such ecosystems would be more vulnerable to ecological and seawater  
47 chemistry changes, impacting local economy.

49 Since AR5, (Ciais et al., 2013) and in agreement with SROCC (IPCC, 2019b), there is now *high agreement*  
50 (*robust evidence*) that coastal ocean acidification, whether induced only by increasing atmospheric CO<sub>2</sub> or  
51 exacerbated by eutrophication or upwelling, has negative effects on specific groups of marine organisms  
52 such as reef-building corals, crabs, pteropods, and sessile fauna (Dupont et al., 2010; Bindoff et al., 2019;  
53 Bednaršek et al., 2020; Osborne et al., 2020), WGII AR6 Chapter 3), especially when combined with  
54 stressors such as temperature and deoxygenation, and potentially increased bioavailability of toxic elements  
55 such as arsenic and copper (Millero et al., 2009; Boyd et al., 2015; Breitburg et al., 2018).

1  
2 Since SROCC (Bindoff et al., 2019), there is further evidence that anthropogenic eutrophication via  
3 continental runoff and atmospheric nutrient deposition, and ocean warming are *very likely* the main drivers  
4 of deoxygenation in coastal areas (Levin and Breitburg, 2015; Levin et al., 2015; Royer et al., 2016;  
5 Breitburg et al., 2018; Cocquempot et al., 2019; Fagundes et al., 2020; Limburg et al., 2020a). Increasing  
6 intensity and frequency of wind-driven upwelling is responsible for longer and more intense coastal hypoxia,  
7 fuelled by organic matter degradation from primary production (*medium to high agreement, medium*  
8 *evidence*) (Rabalais et al., 2010; Bakun et al., 2015; Varela et al., 2015; Fennel and Testa, 2019; Limburg et  
9 al., 2020a). Locally, submarine groundwater discharge may enhance the eutrophication state (*low agreement,*  
10 *limited evidence*, (Luijendijk et al., 2020)). Since AR5 (Ciais et al., 2013) and SROCC (Bindoff et al., 2019)  
11 new observations and model studies confirm the trends in increasing coastal hypoxia caused by  
12 eutrophication, ocean warming and changes in circulation (Claret et al., 2018; Dussin et al., 2019; Limburg  
13 et al., 2020a), as well as the ubiquitous impacts on marine organisms and fisheries (Carstensen and Conley,  
14 2019; Fennel and Testa, 2019; Osma et al., 2020); WGII Chapter 3). Following open ocean deoxygenation  
15 trends, since the 1950s more than 700 coastal regions are being reported as hypoxic (dissolved oxygen  
16 concentration  $<2 \text{ mg O}_2 \text{ L}^{-1}$ ) (Limburg et al., 2020a). Additionally, deoxygenation or increasing severe  
17 hypoxic periods in coastal areas may enhance the sea-to-air fluxes of  $\text{N}_2\text{O}$  and  $\text{CH}_4$  especially through  
18 microbial-mediated processes in the water column-sediment interface (*medium agreement*) (Middelburg and  
19 Levin, 2009; Naqvi et al., 2010b; Farías et al., 2015; Limburg et al., 2020a).

#### 20 21 22 5.3.5.2 *Spatial Characteristics*

23  
24 There is *high agreement (robust evidence)* that heterogeneity implies different responses of coastal regions to  
25 increasing atmospheric  $\text{CO}_2$ , decreasing seawater pH and calcium carbonate saturation state, and  
26 deoxygenation (Duarte et al., 2013; Regnier et al., 2013b; Breitburg et al., 2018; Laruelle et al., 2018;  
27 Carstensen and Duarte, 2019).

28  
29 There is *high agreement* that long time series of observations utilising standard methods are needed to  
30 distinguish the climate change signal in seawater carbonate chemistry from the natural variability of coastal  
31 sites (Duarte et al., 2013; Salisbury and Jönsson, 2018; IOC, 2019; Sutton et al., 2019; Tilbrook et al., 2019;  
32 Turk et al., 2019). Despite increasing availability of data and sea-air  $\text{CO}_2$  flux budgets for the coastal ocean  
33 (Sections 5.3.5.1, 5.2.3.1), additional long-term observations are required to constrain the global time of  
34 emergence of coastal acidification. There is *high agreement (medium evidence)* that, for the coastal  
35 subtropical to temperate northeast Pacific and northwest Atlantic, the mean time of emergence for  
36 acidification is above two decades (Sutton et al., 2019; Turk et al., 2019).

37  
38 Observations and models predict for the northeast Pacific coastal upwelling area an expansion and  
39 intensification of low-pH deep water intrusions (*high agreement, robust evidence*) (Hauri et al., 2013; Feely  
40 et al., 2016b; Cai et al., 2020). There, areas such as the California Current System are naturally exposed to  
41 intrusions of low-pH, high  $\text{pCO}_{2\text{sea}}$  deep waters from remineralisation processes and anthropogenic  $\text{CO}_2$   
42 intrusion (Feely et al., 2008, 2010, 2018; Chan et al., 2019; Lilly et al., 2019; Cai et al., 2020). The eastern  
43 Pacific coastal upwelling displays seasonality in subsurface aragonite undersaturation as a consequence of  
44 the interplay between anthropogenic  $\text{CO}_2$ , respiration and intrusion of upwelling waters (Feely et al., 2008,  
45 2010, 2016, 2018; Hauri et al., 2013; Vargas et al., 2016; Chan et al., 2019; Lilly et al., 2019). The coastal  
46 southeast Pacific, upwelling combined with low-pH, low-alkalinity, organic matter-rich river inputs display  
47 extreme temporal variability in surface seawater  $\text{pCO}_2$  and low aragonite saturation (Vargas et al., 2016;  
48 Osma et al., 2020).

49  
50 Temperate, non-upwelling coastal areas along the northwest Atlantic display a trend of decreasing seawater  
51 pH, mainly attributed to the combined effects of decreasing seawater buffering capacity and eutrophication  
52 (*high agreement, robust evidence*). Observations show an increasing north to south gradient of aragonite  
53 saturation state (Sutton et al., 2016; Fennel et al., 2019; Cai et al., 2020). Low alkalinity and total inorganic  
54 carbon concentration, combined with an ocean signal of acidification, diminishes the buffering capacity  
55 along the decreasing salinity gradient from the ocean to the coast. Models suggest that in this area the



1 aragonite saturation is seasonally controlled by nutrient availability and primary production, supporting that  
2 eutrophication is the main driver for exacerbating acidification.(Cai et al., 2017, 2020). The coastal Gulf of  
3 Mexico is facing a parallel increase in bottom water acidification and deoxygenation off the Mississippi  
4 Delta driven by eutrophication (Cai et al., 2011; Laurent et al., 2017; Fennel et al., 2019).

5  
6 Many coastal tropical are under heavy anthropogenic eutrophication induced by the effluents from large  
7 cities or receive large riverine inputs of freshwater, nutrients, and organic matter (such as Amazon,  
8 Mississippi, Orinoco, Congo, Mekong, or Changjiang rivers). Under strong eutrophication, often sub-surface  
9 and bottom waters present pH values lower than average surface open ocean (about 8.0) because increased  
10 respiration decreases pH (*high agreement, robust evidence*), despite a net atmospheric CO<sub>2</sub> sink in shallow  
11 and vertically stratified coastal areas (Koné et al., 2009; Wallace et al., 2014; Cotovicz Jr. et al., 2015b;  
12 Cotovicz et al., 2018; Fennel and Testa, 2019; Lowe et al., 2019) (Section 5.3.5.1).

13  
14 There is *medium evidence* from observations and models that the coastal northwestern Antarctic Peninsula  
15 (Southern Ocean) will experience calcium carbonate undersaturation by 2060, considering that  
16 anthropogenic emissions reach an atmospheric CO<sub>2</sub> concentration of about 500 pm at that date (Lencina-  
17 Avila et al., 2018; Monteiro et al., 2020a). The synergies among warming, meltwater, sea-air CO<sub>2</sub>  
18 equilibrium and circulation may, to some extent, offset the coastal ocean acidification trends in Antarctica  
19 (Henley et al., 2020). In the coastal western Arctic Ocean, there is increasing *robust evidence* that ocean  
20 acidification is driven by sea-air CO<sub>2</sub> fluxes and sea-ice melt and increasing intrusions since the 1990s of  
21 low-alkalinity Pacific water, lowering aragonite saturation state (Qi et al., 2017, 2020; Cross et al., 2018).  
22 The Bering Sea (northeastern Pacific) shows decreasing trends in calcium carbonate saturation, associated to  
23 the increasing atmospheric CO<sub>2</sub> uptake combined with riverine freshwater and carbon inputs (*high*  
24 *agreement, robust evidence*) (Pilcher et al., 2019; Sun et al., 2020a).

25  
26 The spatial distribution of hypoxic areas is highly heterogeneous in the coastal ocean, and there is *high*  
27 *agreement, robust evidence* that more severe hypoxia or anoxia is often associated to highly populated  
28 coastal areas, or local circulation and upwelling, and seasonal stratification lead to an accumulation of  
29 organic matter in subsurface waters (Ciais et al., 2013; Rabalais et al., 2014; Li et al., 2016b; Breitburg et al.,  
30 2018; Bindoff et al., 2019) (SROCC Chapter 5; IPCC, (2019b)). The causes and trends of coastal  
31 deoxygenation can only be assessed by making available long-term time series combined to regional  
32 modelling (Fennel and Testa, 2019), as in the California current system (Wang et al., 2017), the East China  
33 Sea (Chen et al., 2007; Qian et al., 2017), the Namibian or along the northwestern Atlantic shelves (Claret et  
34 al., 2018). Other coastal upwelling sites such as the Arabian Sea display seasonal hypoxia but no worsening  
35 trends (Gupta et al., 2016).

36  
37 The Baltic Sea is the largest semi-enclosed sea where hypoxia is reported to have happened before the 1950s  
38 (Carstensen et al., 2014; Rabalais et al., 2014; Łukawska-Matuszewska et al., 2019). The frequency and  
39 volume of seawater inflow from the North Sea decreased after 1950, leading to an expansion of hypoxic  
40 areas from 40,000 to 60,000 km<sup>2</sup> in combination with increasing eutrophication (Carstensen et al., 2014).  
41 From the available observations, there is *robust evidence* that many areas in the Baltic Sea are experiencing  
42 deoxygenation despite efforts to reduce nutrient loads (Lennartz et al., 2014; Jokinen et al., 2018).

43  
44 There is *medium agreement (medium evidence)* that just by reducing anthropogenic nutrient inputs may lead  
45 to less severe coastal hypoxic conditions, as observed in the coastal north-western Adriatic Sea (Djakovac et  
46 al. (2015). However, low-oxygen sediments may remain a long-term source of phosphorus and ammonium  
47 to the water column, fuelling primary production (Jokinen et al., 2018; Fennel and Testa, 2019; Limburg et  
48 al., 2020b).

#### 51 **5.4 Biogeochemical Feedbacks on Climate Change**

52  
53 This section covers biogeochemical feedbacks on climate change, which represent one of the largest sources  
54 of uncertainty in climate change projections. The relevant processes are discussed (Sections 5.4.1 to 5.4.4),  
55 prior to discussing the simulation and projection of the carbon cycle in Earth system models (Section 5.4.5),

1 emergent constraints on future projections (Section 5.4.7), non-CO<sub>2</sub> feedbacks (Section 5.4.7), and possible  
2 biogeochemical abrupt changes (Section 5.4.8).

### 5.4.1 Direct CO<sub>2</sub> Effect on Land Carbon Uptake

7 AR5 (WGI, Box 6.3) and SRCCL (IPCC, 2019a) concluded with *high confidence* that rising atmospheric  
8 CO<sub>2</sub> increases leaf-level photosynthesis. This effect is represented in all ESMs. New studies since AR5 add  
9 evidence that the leaf-level CO<sub>2</sub> fertilisation is modulated by acclimation of photosynthesis to long-term CO<sub>2</sub>  
10 exposure, growth temperature, seasonal drought, and nutrient availability, but these effects are not yet  
11 routinely represented in ESMs (Smith and Dukes, 2013; Baig et al., 2015; Kelly et al., 2016; Drake et al.,  
12 2017; Jiang et al., 2020a). Cross-Chapter Box 5.1 assesses multiple lines of evidence, which suggest that the  
13 ratio of plant CO<sub>2</sub> uptake to water loss (plant water-use efficiency; WUE) increases in near proportionality to  
14 atmospheric CO<sub>2</sub>. Despite advances in the regional coverage of field experiments, observations of the  
15 consequences of CO<sub>2</sub> fertilisation at ecosystem level are still scarce, in particular from outside the temperate  
16 zone (Song et al., 2019a). New syntheses since AR5 corroborate that the effect of elevated CO<sub>2</sub> on plant  
17 growth and ecosystem carbon storage is generally positive (*high confidence*), but is modulated by  
18 temperature, water and nutrient availability (Reich et al., 2014; Obermeier et al., 2017; Peñuelas et al., 2017;  
19 Hovenden et al., 2019; Song et al., 2019a). Plant carbon allocation, changes in plant community  
20 composition, disturbance, and natural plant mortality are important processes affecting the magnitude of the  
21 response, but are currently poorly represented in models (De Kauwe et al., 2014; Friend et al., 2014; Reich et  
22 al., 2018; Walker et al., 2019a; Yu et al., 2019), and thus contribute strongly to uncertainty in ESM  
23 projections (Arora et al., 2020).

25 Field studies with elevated CO<sub>2</sub> have demonstrated that the initial stimulation of above-ground growth may  
26 decline if insufficient nutrients such as nitrogen or phosphorus are available (Finzi et al., 2007; Norby et al.,  
27 2010; Hungate et al., 2013; Reich and Hobbie, 2013; Talhelm et al., 2014; Terrer et al., 2018). Model-data  
28 syntheses have demonstrated that the ability to capture the observed long-term effect of elevated CO<sub>2</sub>  
29 depends on the ability of models to predict the effect of vegetation on soil biogeochemistry (Zaehle et al.,  
30 2014; Koven et al., 2015; Medlyn et al., 2015; Walker et al., 2015). Meta-analyses of CO<sub>2</sub> manipulation  
31 experiments point to increased soil microbial activity and accelerated turnover of soil organic matter (van  
32 Groenigen et al., 2017) as a result of increased below-ground carbon allocation by plants (Song et al.,  
33 2019b), and increased root exudation or mycorrhizal activity due to enhanced plant nutrient requirements  
34 under elevated CO<sub>2</sub> (Drake et al., 2011; Terrer et al., 2016; Meier et al., 2017). These effects are not  
35 considered in most ESMs. One global model that attempts to represent these processes suggests that elevated  
36 CO<sub>2</sub> related carbon accumulation is reduced in soils but increased in vegetation relative to more conventional  
37 models (Sulman et al., 2019).

39 Our understanding of the effects of phosphorus limitation is less developed than for nitrogen, but a growing  
40 body of literature suggests it is as important, particularly in regions with highly weathered soils (Wang et al.,  
41 2018; Terrer et al., 2019; Du et al., 2020). CO<sub>2</sub> experiments collectively show that soil phosphorus is an  
42 important constraint on the CO<sub>2</sub> fertilisation effect on plant biomass (Terrer et al., 2019; Jiang et al., 2020a).  
43 Indeed, a free-air CO<sub>2</sub> enrichment experiment in a phosphorus-limited mature forest ecosystem did not find  
44 an increase in biomass production despite increases in photosynthesis after four years of CO<sub>2</sub> exposure  
45 (Jiang et al., 2020b). The lack of free-air CO<sub>2</sub> enrichment experiments in phosphorus-limited tropical forests  
46 limits our understanding on the role of phosphorus availability in constraining the CO<sub>2</sub> fertilisation effect  
47 globally (Norby et al., 2016; Fleischer et al., 2019). Models accounting for the effects of phosphorus  
48 availability, in addition to nitrogen, generally show an even stronger reduction of the response of ecosystem  
49 carbon storage to elevated CO<sub>2</sub> (Goll et al., 2012; Zhang et al., 2014; Yang et al., 2019b). Insufficient data  
50 and uncertainties in the process formulation cause large uncertainty in the magnitude of this effect (Medlyn  
51 et al., 2016; Fleischer et al., 2019).

53 Consistent with AR5 (WGI, Section 6.4.2), the CO<sub>2</sub> fertilisation effect is the dominant cause for the projected  
54 increase in land carbon uptake between 1860 and 2100 in ESMs (Figures 5.26, 5.27; Table 5.5; Arora et al.,  
55 2020). In the CMIP6 ensemble, the increase of land carbon storage due to CO<sub>2</sub> fertilisation is a global

1 phenomenon but is strongest in the tropics (Figure 5.26). The resulting increase of productivity is a key  
2 driver of increases in vegetation and soil carbon storage. However, consistent with earlier findings (Todd-  
3 Brown et al., 2013; Friend et al., 2014; Hajima et al., 2014), processes affecting vegetation carbon-use  
4 efficiency and turnover such as allocation changes, mortality, and vegetation structural changes, as well as  
5 the pre-industrial soil carbon turnover time, also play an important role (Arora et al., 2020).

6  
7 As a major advance since AR5 (WGI, Section 6.4.2), 6 out of 11 models in the C<sup>4</sup>MIP-CMIP6 ensemble  
8 account for nitrogen cycle dynamics over land (Table 5.4). On average, these models exhibit a 25–30%  
9 lower CO<sub>2</sub> fertilisation effect on land carbon storage, compared to models that do not (Figures 5.29, Table  
10 5.5). The only model in the C<sup>4</sup>MIP-CMIP6 ensemble that explicitly represents the effect of P availability on  
11 plant growth suggests the lowest C storage response to increasing CO<sub>2</sub> (Arora et al., 2020). The lower CO<sub>2</sub>  
12 effect due to decreased nutrient availability is generally consistent with analyses of the implicit nutrient  
13 limitation in CMIP5 simulations (Wieder et al., 2015; Zaehle et al., 2015) and independent assessments by  
14 stand-alone land models (Zaehle et al., 2010; Wårlind et al., 2014; Zhang et al., 2014; Goll et al., 2017;  
15 Meyerholt et al., 2020a). The simulated effects are generally consistent with expectations based on  
16 independent observations (Walker et al., 2020). However, the magnitude of nutrient feedbacks in these  
17 models is poorly constrained by observations, owing to the limited geographic distribution of available  
18 observations and the uncertain scaling of results obtained from manipulation experiments to transient system  
19 dynamics (Song et al., 2019a; Wieder et al., 2019; Meyerholt et al., 2020a).

20  
21 Our understanding of the various biological processes, which affect the strength of the CO<sub>2</sub> fertilisation  
22 effect on photosynthesis and its impact on carbon storage in vegetation and soils, in particular regarding the  
23 limitations imposed by nitrogen and phosphorus availability, has developed since AR5 (WGI, Box 6.2).  
24 Based on consistent behaviour across all CMIP6 ESMs, there is *high confidence* that CO<sub>2</sub> fertilisation of  
25 photosynthesis acts as an important negative feedback on anthropogenic climate change, by reducing the rate  
26 at which CO<sub>2</sub> accumulates in the atmosphere. Since AR5 (WGI, Box 6.2), an increasing number of CMIP6  
27 ESMs account for nutrient cycles and the consistent results found in their model projections suggests with  
28 *high confidence* that limited nutrient availability will limit the CO<sub>2</sub> fertilisation effect (Arora et al., 2020).  
29 The magnitude of both the direct CO<sub>2</sub> effect on land carbon uptake, and its limitation by nutrients, remains  
30 uncertain (*low confidence*).

#### 31 32 33 **5.4.2 Direct CO<sub>2</sub> Effects on Projected Ocean Carbon Uptake**

34  
35 In AR5 (WGI, Section 6.4.2) there was *high agreement* that CMIP5 ESMs project continued ocean CO<sub>2</sub>  
36 uptake through to 2100, with higher uptake corresponding to higher concentration or emission pathways.  
37 There has been no significant change in the magnitude of the sensitivity of ocean carbon uptake to increasing  
38 atmospheric CO<sub>2</sub>, or in the inter-model spread, between the CMIP5 and CMIP6 era (Arora et al., 2020). The  
39 analysis from both emissions and concentration driven CMIP5 models projections show that the ocean sink  
40 stops growing beyond 2050 across all emission scenarios (Section 5.4.5.3). CMIP6 models also show a  
41 similar time-evolution of global ocean CO<sub>2</sub> uptake to CMIP5 models over the 21st century (Figure 5.25)  
42 with decreasing net ocean CO<sub>2</sub> uptake ratio to anthropogenic CO<sub>2</sub> emissions under SSP5–8.5.

43  
44 The projected weakening of ocean carbon uptake is driven by a combination of decreasing carbonate  
45 buffering capacity and warming which are positive feedbacks under weak to no mitigation scenarios (SSP4  
46 and 5). In high mitigation scenarios (SSP1–2.6), weakening ocean carbon uptake is driven by decreasing  
47 emissions (Cross-Chapter Box 5.3). The detailed understanding of carbonate chemistry in seawater that has  
48 accumulated over more than half a century (e.g. Egleston et al., 2010; Revelle & Suess, 1957), provides *high*  
49 *confidence* that the excess CO<sub>2</sub> dissolved in seawater leads to a nonlinear reduction of the CO<sub>2</sub> buffering  
50 capacity, that is smaller dissolved inorganic carbon (DIC) increase with respect to pCO<sub>2</sub> increase along with  
51 the increase in cumulative ocean CO<sub>2</sub> uptake. Recent studies (Katavouta et al., 2018; Jiang et al., 2019;  
52 Arora et al., 2020; Rodgers et al., 2020) suggest with *medium confidence* that the decrease in the ocean CO<sub>2</sub>  
53 uptake ratio to anthropogenic CO<sub>2</sub> emission under low to no mitigation scenarios over the 21st century is  
54 predominantly attributable to the ocean carbon-concentration feedback through the reduction of the seawater  
55 CO<sub>2</sub> buffering capacity but with contributions from physical drivers such as warming and wind stress

1 (*medium confidence*) and biological drivers (*low confidence*) (Sections 5.2.1.3.3, 5.4.4).

2  
3 Projected increases in ocean DIC due to anthropogenic CO<sub>2</sub> uptake amplify the sensitivity of carbonate  
4 system variables to perturbations of DIC in the surface ocean, for example via the amplitude of the seasonal  
5 cycle of pCO<sub>2</sub>, which impact the mean annual air-sea fluxes (Fassbender et al., 2018; Hauck et al., 2015;  
6 Landschützer et al., 2018; SROCC Section 5.2.2.3). A larger amplification of the surface ocean pCO<sub>2</sub>  
7 seasonality occurs in the subtropics where pCO<sub>2</sub> seasonality is dominated by temperature seasonality, with  
8 the summer increase in the difference in pCO<sub>2</sub> between surface water and the overlying atmosphere reaching  
9 3µatm per decade between 1990 and 2030 under RCP8.5 (Schlunegger et al., 2019; Rodgers et al., 2020). In  
10 contrast, the impact of biological production on the seasonal cycle of pCO<sub>2</sub> in summer in the Southern Ocean  
11 strengthens the drawdown of CO<sub>2</sub> (Hauck et al., 2015).

12  
13 Overall, there is *medium confidence* on three outcomes in the ocean from projected CO<sub>2</sub> uptake under  
14 medium to high CO<sub>2</sub> concentration scenarios: (i) a strengthening positive feedback, which impacts on the  
15 airborne fraction via the reduction of the ocean CO<sub>2</sub> buffering capacity due to cumulative ocean CO<sub>2</sub> uptake,  
16 which reduces the net ocean CO<sub>2</sub> uptake ratio to anthropogenic CO<sub>2</sub> emission (Katavouta et al., 2018; Arora  
17 et al., 2020; Rodgers et al., 2020); (ii) an amplification of the seasonal cycle of CO<sub>2</sub> variables, which impact  
18 both the ocean sink and ocean acidification (Hauck et al., 2015); (iii) a decrease in the aragonite and calcite  
19 saturation levels in the ocean which negatively impacts the calcification rates of marine organisms (*high*  
20 *confidence*) and which forms a negative feedback on the uptake of CO<sub>2</sub> (McNeil and Sasse, 2016) (Cross-  
21 Chapter Box 5.3).

### 22 23 24 **5.4.3 Climate Effect on Land Carbon Uptake**

25  
26 AR5 assessed with *medium confidence*, that future climate change will decrease land carbon uptake relative  
27 to the case with constant climate, but with a poorly-constrained magnitude (WGI, Chapter 6, Executive  
28 Summary). Ongoing uncertainty in the magnitude and geographic pattern of the feedbacks (Section 5.4.5),  
29 continues to support a *medium confidence* assessment that future climate change will decrease land carbon  
30 uptake relative to the case with constant climate.

#### 31 32 33 **5.4.3.1 Plant Physiology**

34  
35 Plant productivity is highly dependent on local climate. In cold environments, warming has generally led to  
36 an earlier onset of the growing season, and with it an increase in early-season vegetation productivity (e.g.  
37 Forkel et al., 2016). However, this trend is affected by adverse effects of climate variability, and other  
38 emerging limitations on vegetation production by water, energy and nutrients, which may gradually reduce  
39 the effects of warming (Piao et al., 2017; Buermann et al., 2018; Liu et al., 2019). At centennial timescales,  
40 boreal forest expansion may act as a climate-driven carbon sink (Pugh et al., 2018).

41  
42 In tropical and temperate environments, temperature simultaneously affects the metabolic rates of  
43 photosynthetic processes within leaf tissues as well as the vapor pressure deficit that drives transpiration, its  
44 control by leaf stomata, and the resulting soil and plant tissue water content. Thus the direct effect of  
45 warming on photosynthesis can be positive, negative, or invariant depending on environmental context (Lin  
46 et al., 2012; Yamori et al., 2014; Smith and Dukes, 2017; Grossiord et al., 2020). Observations and models  
47 suggest that the vapour pressure deficit effects are stronger than direct temperature effects on enzyme  
48 activities (Smith et al., 2020), and that acclimation of photosynthetic optimal temperature may mitigate  
49 productivity losses of tropical forests under climate change (Kattge and Knorr, 2007; Tan et al., 2017;  
50 Kumarathunge et al., 2019). Some models have begun to include these acclimation responses, both in  
51 photosynthesis and autotrophic respiration (Lombardozzi et al., 2015; Smith et al., 2015; Huntingford et al.,  
52 2017; Mercado et al., 2018).

#### 53 54 55 **5.4.3.2 Fire and Other Disturbances**

1  
2 SRCCL assessed that climate change is playing an increasing role in determining wildfire regimes alongside  
3 human activity (*medium confidence*), with future climate variability expected to enhance the recurrence and  
4 severity of wildfires in many biomes, such as tropical rainforests (*high confidence*). Projections of increased  
5 fire weather in a warmer climate are widespread (Section 12.3.2.8) and may drive increased fire frequency  
6 and severity in several regions, including Arctic and boreal ecosystems (Gauthier et al., 2015; Walker et al.,  
7 2019b), Mediterranean-type ecosystems (Turco et al., 2014; Jin et al., 2015), degraded tropical forests  
8 (Aragão et al., 2018), and tropical forest-savanna transition zones (Lehmann et al., 2014).

9  
10 Wildfire is included in some CMIP6 ESMs (Table 5.4) and is thus only partially represented in estimates of  
11 carbon-climate feedbacks from these models. The CMIP5 ESMs that include fire project 8–58% increases of  
12 fire carbon emissions under future scenarios, with higher emissions under higher-warming scenarios; the  
13 ensemble spread is driven by differing factors such as population density, fire management, and other land-  
14 use processes (Kloster and Lasslop, 2017). Fire dynamics in CMIP6 models, as evaluated in land-only  
15 configurations of CMIP6-generation land surface models, also show large variations but better agreement  
16 with observations (Teckentrup et al., 2019; Hantson et al., 2020; Lasslop et al., 2020).

17  
18 Climate change also drives changes to vegetation composition and ecosystem carbon storage through other  
19 disturbances such as forest dieback that lead to biome shifts in tropical forests (Cox et al., 2004; Jones et al.,  
20 2009; Brando et al., 2014; Le Page et al., 2017; Zemp et al., 2017), and temperate and boreal regions (Joos et  
21 al., 2001; Lucht et al., 2006; Scheffer et al., 2012; Lasslop et al., 2016). AR5 assessed that large-scale loss of  
22 tropical forests due to climate change is *unlikely* (WGI, Section 6.4.9). Newer ecosystem modelling  
23 approaches that include a greater degree of ecosystem heterogeneity and diversity show a reduced sensitivity  
24 of such forest dieback-type changes (Levine et al., 2016; Sakschewski et al., 2016), supporting the AR5  
25 assessment (Section 5.4.8). Beyond such biome shifts, observations of tropical forests also show that  
26 increasing tree mortality rates within tropical forests may reduce carbon turnover times and storage (Brienen  
27 et al., 2015), that increased tree mortality rates in tropical forests and elsewhere are expected with increased  
28 temperatures and vapor pressure deficit (Cross-chapter Box 5.1; (Allen et al., 2015; McDowell et al., 2018;  
29 Grossiord et al., 2020), and that these processes are not well represented in ESMs (Powell et al., 2013; Fisher  
30 et al., 2018). An ensemble of land models that include ecological processes such as forest demography  
31 shows that changes to mortality may be a more important driver of carbon dynamics than changes to  
32 productivity (Friend et al., 2014).

33  
34 Overall, climate change will force widespread increases in fire weather throughout the world (Section  
35 12.3.2.8). Because of incomplete inclusion of fire in ESMs, a separate compilation of fire-driven carbon-  
36 climate feedback estimates (Eliseev et al., 2014a; Harrison et al., 2018) (section 5.4.8). There is *low*  
37 *agreement* in magnitude and *medium agreement* in sign, which alongside other literature (Jones et al., 2020),  
38 leads to an assessment of *medium confidence* that fire represents a positive carbon-climate feedback, but *very*  
39 *low confidence* in the magnitude of that feedback. Other disturbances such as tree mortality will increase  
40 across several ecosystems (*medium agreement*) with decreased vegetation carbon (*medium confidence*).  
41 However, the lack of model agreement and lack of key process representation in ESMs lead to a *low*  
42 *confidence* assessment in the projected magnitude of this feedback.

### 43 44 45 5.4.3.3 Soil Carbon

46  
47 Changes to soil carbon stocks in response to climate change are a potentially strong positive feedback (Cox  
48 et al., 2000). Since the AR5 (WGI, Section 6.4.2), there has been progress made in understanding soil carbon  
49 dynamics, and associated feedbacks. These include: (i) an increased understanding of and ability to quantify  
50 high latitude soil carbon feedbacks (Box 5.1); (ii) increased understanding of the causes responsible for soil  
51 carbon persistence on long timescales, particularly the interactions between decomposers and soil organic  
52 matter and mineral assemblages (Kleber et al., 2007; Schmidt et al., 2011; Luo et al., 2016); and (iii)  
53 increased understanding of soil carbon dynamics in subsurface layers (Hicks Pries et al., 2017; Balesdent et  
54 al., 2018).

1 CMIP6 ESMs predict losses of soil carbon with warming, which are larger than climate-driven vegetation  
2 carbon losses (Arora et al., 2020). As in CMIP5 (Todd-Brown et al., 2013), there is also a large CMIP6  
3 ensemble spread in climate-driven soil carbon changes, partially driven by a large spread in the current soil  
4 carbon stocks predicted by the models. In CMIP5 ESMs, much of the soil carbon losses with warming can be  
5 traced to decreased carbon inputs, with a weaker contribution from changing soil carbon lifetimes due to  
6 faster decomposition rates (Koven et al., 2015a), which may be an artefact of the lack of permafrost carbon  
7 (Box 5.1). Isotopic constraints suggest that CMIP5 ESMs systematically overestimated the transient  
8 sensitivity of soil <sup>14</sup>C responses to atmospheric <sup>14</sup>C changes, implying that the models respond too quickly to  
9 changes in either inputs or turnover times and that the soil contribution to all feedbacks may thus be weaker  
10 than currently projected (He et al., 2016). Using natural gradients of soil carbon turnover as a constraint on  
11 long-term responses to warming suggests that both CMIP5 and CMIP6 ESMs may systematically  
12 underestimate the temperature sensitivity at high latitudes, and may overestimate the temperature sensitivity  
13 in the tropics (Koven et al., 2017; Wieder et al., 2018; Varney et al., 2020), although experimental soil  
14 warming in tropical forests suggest high sensitivity of decomposition to warming in those regions as well  
15 (Nottingham et al., 2020).

16  
17 Peat soils, where thick organic layers build up due to saturated and anoxic conditions, represent another  
18 possible source of carbon to the atmosphere. Peats could dry, and decompose or burn as a result of climate  
19 change in both high (Chaudhary et al., 2020) and tropical (Cobb et al., 2017) latitudes, and in combination  
20 with anthropogenic drainage of peatlands (Warren et al., 2017). Peat carbon dynamics are not included in the  
21 majority of CMIP6 ESMs.

22  
23 Soil microbial dynamics shift in response to temperature, giving rise to complex longer-term trophic effects  
24 that are more complex than the short-term sensitivity of decomposition to temperature. Such responses are  
25 observed in response to long-term warming experiments (Melillo et al., 2017). While most CMIP6 ESMs do  
26 not include microbial dynamics, simplified global soil models that do include such dynamics show greater  
27 uncertainty in projections of soil C changes, despite agreeing more closely with current observations, than  
28 the linear models used in most ESMs (Wieder et al., 2013; Guenet et al., 2018).

29  
30 In nutrient limited ecosystems, prolonged soil warming can induce a fertilisation effect through increased  
31 decomposition, which increases nutrient availability and thereby vegetation productivity (Melillo et al.,  
32 2011). Models that include this process tend to show a weaker carbon-climate feedback than those that do  
33 not (Thornton et al., 2009; Zaehle et al., 2010; Wårlind et al., 2014; Meyerholt et al., 2020b). In CMIP6, 6  
34 out of 11 ESMs include a representation of the nitrogen cycle, and the mean of those models predicts a  
35 weaker carbon-climate feedback than the overall ensemble mean (Arora et al., 2020) (Section 5.4.8). These  
36 models only partly account for the interactions of nutrient effects with other processes such as shifts of  
37 vegetation zones under climate changes (Sakaguchi et al., 2016) leading to either changes in species  
38 composition or changes in plant tissue nutrient to carbon ratios (Thomas et al., 2015; Achat et al., 2016; Du  
39 et al., 2019).

40  
41 The *high agreement* and multiple lines of evidence that warming increases decomposition rates lead to *high*  
42 *confidence* that warming will overall result in carbon losses relative to a constant climate and contribute to  
43 the positive carbon-climate feedback (Section 5.4.8). However, the widespread in ESM projections and lack  
44 of model representation of key processes that may amplify or mitigate soil carbon losses on longer  
45 timescales (including microbial dynamics, permafrost, peatlands, and nutrients) lead to *low confidence* in the  
46 magnitude of global soil carbon losses with warming.

47  
48  
49 **[START BOX 5.1 HERE]**

#### 50 **BOX 5.1: Permafrost Carbon and Feedbacks to Climate**

##### 51 **What is permafrost carbon and why should we be concerned about it?**

52  
53  
54  
55 Soils in the Arctic and other cold regions contain perennially frozen layers, known as permafrost. Soils in the

1 northern permafrost region store a large amount of organic carbon, estimated at 1460–1600 PgC across both  
2 surface soils and deeper deposits (Hugelius et al., 2014; Strauss et al., 2017; Mishra et al., 2021). Of that  
3 carbon, permafrost soils and deposits store 1070–1360 PgC, of which 300–400 PgC are in the first meter,  
4 and the rest at depth. The remaining 280–340 Pg C are in permafrost-free soils within the permafrost region.  
5 These carbon deposits have accumulated over thousands of years due to the slow rates of organic matter  
6 decomposition in frozen and/or waterlogged soil layers, but these frozen soils are highly decomposable upon  
7 thaw (Schädel et al., 2014).

### 9 **Is permafrost carbon already thawing and emitting greenhouse gases?**

10  
11 The permafrost region was a historic carbon sink over centuries to millennia (*high confidence*) (Loisel et al.,  
12 2014; Lindgren et al., 2018). Currently though, thawing soils due to anthropogenic warming are losing  
13 carbon from the decomposition of old frozen organic matter, as found via  $^{14}\text{C}$  signature of respiration at sites  
14 undergoing rapid permafrost thaw (Hicks Pries et al., 2013), of dissolved organic carbon in rivers draining  
15 watersheds with permafrost thaw (Vonk et al., 2015; Wild et al., 2019), and of  $\text{CH}_4$  produced in thawing  
16 lakes (Walter Anthony et al., 2016).

17  
18 Despite accumulating evidence of increased carbon losses, it is difficult to scale up site- and ecosystem-level  
19 measurements to assess the net carbon balance over the entire permafrost region, due to the high spatial  
20 heterogeneity, the strong seasonal cycles and the difficulty in monitoring these regions consistently across  
21 the year. SROCC assessed with *high confidence* both that ecosystems in the permafrost region act as carbon  
22 sinks during the summer growing season, and that wintertime carbon losses are significant, consistent with a  
23 multi-decadal small increase in  $\text{CO}_2$  emissions during early winter at Barrow, Alaska (Sweeney et al., 2016;  
24 Webb et al., 2016; Meredith et al., 2019). These findings have been further strengthened by recent  
25 comprehensive synthesis of in-situ wintertime flux observations that show large carbon losses during the  
26 non-growing season (Natali et al., 2019). Increased autumn and winter respiration are a key large-scale  
27 fingerprint of top-down permafrost thaw predicted by ecosystem models (Parazoo et al., 2018). However, the  
28 length of these wintertime observational records is too short to unequivocally determine whether winter  
29 carbon losses are higher now than they used to be. One study inferred a multi-year net  $\text{CO}_2$  source for the  
30 tundra in Alaska (Commane et al., 2017), which is equivalent to  $0.3 \text{ PgC yr}^{-1}$  when scaled up to the northern  
31 permafrost region (*low confidence*) (Meredith et al., 2019).

32  
33 Since the AR5, evidence of a more active carbon cycle in the northern high latitude regions has also been  
34 observed through the increased amplitude of  $\text{CO}_2$  seasonal cycles. However, the relative roles of local  
35 sources versus influence from mid-latitudes makes it difficult to infer changes to Arctic ecosystems from  
36 these observations (Graven et al., 2013; Forkel et al., 2016; Takata et al., 2017; Bruhwiler et al., 2021).  
37 Estimates of  $\text{CO}_2$  fluxes with atmospheric inversion models showed an enhanced seasonal cycle amplitude  
38 but no significant trends in annual total fluxes, in agreement with flux tower measurements over one decade  
39 (2004–2013) (Welp et al., 2016; Takata et al., 2017).

40  
41 In addition to  $\text{CO}_2$ ,  $\text{CH}_4$  emissions from the northern permafrost region contribute to the global methane  
42 budget, but evidence as to whether these emissions have increased from thawing permafrost is mixed.  
43 SROCC assigned *low confidence* to the degree of recent additional methane emissions from diverse sources  
44 throughout the permafrost region. These include observed regional lake area change, which suggest a  $1.6\text{--}5$   
45  $\text{Tg CH}_4 \text{ yr}^{-1}$  increase over the last 50 years (Walter Anthony et al., 2016), ice-capped geological sources  
46 (Walter Anthony et al., 2012; Kohnert et al., 2017), and shallow Arctic Ocean shelves. The shallow subsea  
47 emissions are particularly uncertain due to both the wide range of estimates ( $3 \text{ Tg CH}_4 \text{ yr}^{-1}$  (Thornton et al.,  
48 2016a) to  $17 \text{ Tg CH}_4 \text{ yr}^{-1}$  (Shakhova et al., 2014)), and the lack of a baseline with which to infer any  
49 changes; the upper half of this range in flux estimates is, however, inconsistent with the atmospheric  
50 inversions constrained by the pan-Arctic  $\text{CH}_4$  concentration measurements (Berchet et al., 2016).

51  
52 Atmospheric measurements and inversions performed at the global and regional scales do not show any  
53 detectable trends in annual mean  $\text{CH}_4$  emissions from the permafrost region over the past 30 years (Jackson  
54 et al., 2020; Saunio et al., 2020; Bruhwiler et al., 2021), consistent with atmospheric measurements in  
55 Alaska that showed no significant annual trends despite significant increase in air temperature (Sweeney et

1 al., 2016). Atmospheric inversions and biospheric models do not show any clear trends in CH<sub>4</sub> emission for  
2 wetland regions of the high latitudes during the period 2000–2016 (Patra et al., 2016; Poulter et al., 2017;  
3 Jackson et al., 2020; Saunio et al., 2020). Large uncertainties on wetland extent and limited data constraints  
4 place *low confidence* in these modeling approaches.

5  
6 SROCC also assessed with *high confidence* that methane fluxes have been under-observed due to their high  
7 variability at multiple scales in both space and time, and that there is a persistent mismatch between top-  
8 down and bottom-up methane budgets, with emissions calculated by upscaling ground observations typically  
9 higher than emissions inferred from large-scale atmospheric observations (Thornton et al., 2016b).

10  
11 In conclusion, there is *high confidence* that the permafrost region has acted as a historic carbon sink over  
12 centuries to millennia and *high confidence* that some permafrost regions are currently net sources of CO<sub>2</sub>.  
13 There is *robust evidence* that some CH<sub>4</sub> emissions sources and for some regions have increased over the past  
14 decades (*medium confidence*). For the northern permafrost-wide region, no multi-decadal trend has been  
15 detected on CO<sub>2</sub> and CH<sub>4</sub> fluxes but given the low resolution and sparse observations of current observations  
16 and modeling systems, we place *low confidence* in this statement.

17  
18 Since AR5, there have been new studies showing that permafrost thaw also leads to N<sub>2</sub>O release from soil  
19 (Abbott and Jones, 2015; Karelin et al., 2017; Wilkerson et al., 2019), a previously unaccounted source.  
20 However, this release is unquantified at the pan-Arctic scale.

### 21 22 **What does the paleo record tell us about how much emissions to expect?**

23  
24 Large areas of Alaska and Siberia are underlain by frozen, glacial-age, ice- and carbon-rich deposits, and  
25 many of these areas show evidence of thermokarst processes during Holocene warm periods. Rapid warming  
26 of high northern latitudes contributed to permafrost thaw, liberating labile organic carbon to the atmosphere  
27 (Köhler et al., 2014; Crichton et al., 2016; Winterfeld et al., 2018; Meyer et al., 2019), supporting the  
28 vulnerability of these areas to further warming (Strauss et al., 2013, 2017).

29  
30 Radiogenic and stable isotopic measurements on CH<sub>4</sub> trapped in Antarctic ice support the view that CH<sub>4</sub>  
31 emissions from fossil carbon reservoirs, including permafrost and methane hydrates, in response to the  
32 deglacial warming remained small. Mass-balance calculations reveal that geological CH<sub>4</sub> emissions have not  
33 exceeded 19 Tg yr<sup>-1</sup>, highlighting that the deglacial increase in CH<sub>4</sub> emissions were predominantly related to  
34 contemporary methane emissions from tropical wetlands and seasonally inundated floodplains (Bock et al.,  
35 2017; Petrenko et al., 2017; Dyonisius et al., 2020). Isotopic constraints on CO<sub>2</sub> losses from permafrost with  
36 warming after the Last Glacial Maximum (LGM) are weaker than for CH<sub>4</sub>. While the biosphere as a whole  
37 held less carbon during LGM than the preindustrial, that change in stocks was smaller than the change in  
38 plant productivity, and so carbon losses at high latitudes may have been offset by increased tropical  
39 productivity in response to warming during the last deglacial transition (LDT) (Ciais et al., 2012b). There is  
40 also paleoclimate evidence for processes that mitigate carbon losses with warming on longer timescales,  
41 such as longer-term carbon accumulation in lake deposits following thermokarst thaw (Walter Anthony et  
42 al., 2014), and long-term accumulation of carbon in permafrost soils following LDT carbon loss (Lindgren et  
43 al., 2018), particularly in peatlands which accumulated carbon at a slow but persistent rate in warm  
44 paleoclimates (Treat et al., 2019).

45  
46 In conclusion, several independent lines of evidence indicate that permafrost thaw did not release vast  
47 quantities of fossil CH<sub>4</sub> associated with the transient warming events of the LDT, suggesting that large  
48 emissions of CH<sub>4</sub> from old carbon sources will not occur in response to future warming (*medium*  
49 *confidence*).

### 50 51 **How much emissions do we expect in the future?**

52  
53 Near-surface permafrost is projected to decrease significantly under future global warming scenarios (*high*  
54 *confidence*, Section 9.5.2), thus creating the potential for releasing CO<sub>2</sub> and CH<sub>4</sub> to the atmosphere, and act  
55 as a positive carbon-climate feedback.



1  
2 The processes that govern permafrost carbon loss are grouped into gradual and abrupt mechanisms. Gradual  
3 processes include the deepening of the seasonally-thawed active layer into perennially-frozen permafrost  
4 layers and lengthening of the thawed season within the active layer, which increases the amount of organic  
5 carbon that is thawed and the duration of thaw. Abrupt thaw processes include ice-wedge polygon  
6 degradation, hillslope collapse, thermokarst lake expansion and draining, all of which are processes largely  
7 occurring in regions with very high soil carbon content (Olefeldt et al., 2016a, 2016b). Abrupt thaw  
8 processes can contribute up to half of the total net greenhouse gas release from permafrost loss, the rest  
9 attributed to gradual thaw (Schneider von Deimling et al., 2015; Turetsky et al., 2020). Increased fire  
10 frequency and severity (Hu et al., 2010) also contributes to abrupt emissions and the removal of the  
11 insulating cover which leads to an acceleration of permafrost thaw (Genet et al., 2013). Ecological  
12 feedbacks can both mitigate and amplify carbon losses: nutrient release from increased organic matter  
13 decomposition can drive vegetation growth that partially offsets soil carbon losses (Salmon et al., 2016), but  
14 also lead to biophysical feedbacks that further amplify warming (Myers-Smith et al., 2011).

15  
16 Through CMIP5, Earth system models (ESMs) had not included permafrost carbon dynamics. This remains  
17 largely true in CMIP6, with most models not representing permafrost carbon processes, a small number  
18 representing the active-layer thickening effect on decomposition (Table 5.4), and no ESMs representing  
19 thermokarst or fire-permafrost-carbon interactions. The CMIP6 ensemble mean predicts a negative carbon-  
20 climate feedback in the permafrost region. However, those that do include permafrost carbon show a positive  
21 carbon-climate feedback in the permafrost region (Figure 5.27). Given the current limited ESM capacity to  
22 assess permafrost feedbacks, estimates in this report are based on published permafrost-enabled land surface  
23 model results.

24  
25 SROCC assessed that warming under a high emission scenario (RCP8.5 or similar) would result in a loss of  
26 permafrost carbon by 2100 of 10s to 100s of PgC, with a maximum estimate of 240 PgC and a best estimate  
27 of  $92 \pm 17$  PgC (Meredith et al., 2019) (SROCC, Figure 3.11). Under lower emissions scenarios, Schneider  
28 von Deimling et al., (2015) estimated permafrost feedbacks of 20–58 PgC of CO<sub>2</sub> by 2100 under a RCP2.6  
29 scenario, and 28–92 PgC of CO<sub>2</sub> under a RCP4.5 scenario.

30  
31 This new assessment, based on studies included in or published since SROCC (Schaefer et al., 2014; Koven  
32 et al., 2015c; Schneider von Deimling et al., 2015; Schuur et al., 2015; MacDougall and Knutti, 2016a;  
33 Gasser et al., 2018; Yokohata et al., 2020), estimates that the permafrost CO<sub>2</sub> feedback per degree of global  
34 warming (Figure 5.29) is 18 (3.1–41, 5<sup>th</sup>–95<sup>th</sup> percentile range) PgC °C<sup>-1</sup>. The assessment is based on a wide  
35 range of scenarios evaluated at 2100, and an assessed estimate of the permafrost CH<sub>4</sub>-climate feedback at 2.8  
36 (0.7–7.3 5<sup>th</sup>–95<sup>th</sup> percentile range) Pg Ceq °C<sup>-1</sup> (Figure 5.29). This feedback affects the remaining carbon  
37 budgets for climate stabilisation and is included in their assessment (Section 5.5.2).

38  
39 Beyond 2100, models suggest that the magnitude of the permafrost carbon feedback strengthens  
40 considerably over the period 2100–2300 under a high-emissions scenario (Schneider von Deimling et al.,  
41 2015; McGuire et al., 2018). Schneider von Deimling et al., (2015) estimated that thawing permafrost could  
42 release 20–40 PgC of CO<sub>2</sub> in the period from 2100 to 2300 under a RCP2.6 scenario, and 115–172 PgC of  
43 CO<sub>2</sub> under a RCP8.5 scenario. The multi-model ensemble in (McGuire et al., 2018) project a much wider  
44 range of permafrost soil carbon losses of 81–642 PgC (mean 314 PgC) for an RCP8.5 scenario from 2100 to  
45 2300, and of a gain of 14 PgC to a loss of 54 PgC (mean loss of 17 PgC) for an RCP4.5 scenario over the  
46 same period.

47  
48 Methane release from permafrost thaw (including abrupt thaw) under high-warming RCP8.5 scenario has  
49 been estimated at 836–2614 Tg CH<sub>4</sub> over the 21st century and 2800–7400 Tg CH<sub>4</sub> from 2100–2300  
50 (Schneider von Deimling et al., 2015), and as 5300 Tg CH<sub>4</sub> over the 21st century and 16000 Tg CH<sub>4</sub> from  
51 2100–2300 (Turetsky et al., 2020). For RCP4.5, these numbers are 538–2356 Tg CH<sub>4</sub> until 2100 and 2000-  
52 6100 Tg CH<sub>4</sub> from 2100–2300 (Schneider von Deimling et al., 2015), and 4100 Tg CH<sub>4</sub> until 2100 and  
53 10000 Tg CH<sub>4</sub> from 2100–2300 (Turetsky et al., 2020).

54  
55 A key uncertainty is whether permafrost carbon feedbacks scale roughly linearly with warming (Koven et

al., 2015c), or instead scale at a greater (MacDougall and Knutti, 2016b; McGuire et al., 2018) or smaller rate (e.g. CH<sub>4</sub> emissions estimated by Turetsky et al., (2020a)). It also remains unclear whether the permafrost carbon pool represents a coherent global tipping element of the Earth system with a single abrupt threshold (Driijfhout et al., 2015) at a given level of global warming, or a local scale tipping point without abrupt thresholds when aggregated across the pan-Arctic region as is suggested by recent model results (e.g. (Koven et al., 2015b; McGuire et al., 2018)).

In conclusion, thawing terrestrial permafrost will lead to carbon release under a warmer world (*high confidence*). However, there is *low confidence* on the timing, magnitude and linearity of the permafrost climate feedback owing to the wide range of published estimates and the incomplete knowledge and representation in models of drivers and relationships. It is projected that CO<sub>2</sub> released from permafrost will be 18 (3.1–41) PgC per 1°C by 2100 with the relative contribution of CO<sub>2</sub> vs CH<sub>4</sub> remaining poorly constrained. Permafrost carbon feedbacks are included among the underrepresented feedbacks quantified in Figure 5.29.

[END BOX 5.1 HERE]

#### 5.4.4 Climate Effects on Future Ocean Carbon Uptake

##### 5.4.4.1 Physical Drivers of Future Ocean Carbon Uptake and Storage

The principal contribution to increasing global ocean carbon is the air-sea flux of CO<sub>2</sub>, which changes the DIC inventory (Section 5.4.2) (Arora et al., 2020). The processes that influence the variability and trends of the ocean carbon-heat nexus are assessed in Cross-Chapter Box 5.3. Climate has three important impacts on the ocean uptake of anthropogenic CO<sub>2</sub>: (i) ocean warming reduces the solubility of CO<sub>2</sub>, which increases *p*CO<sub>2</sub> and increases the stratification of the mixed layer with both acting as positive feedbacks weakening the ocean sink (Arora et al., 2020; Cross-Chapter Box 5.3; 9.2.1); (ii) changing the temporal and spatial characteristics of wind stress and storm alters mixing – entrainment in the mixed layer and across the bottom of the mixed layer (Bronse laer et al., 2018) and (iii) both warming and wind stress influence the large scale MOC circulation which modifies the rate of ventilation, storage or outgassing of ocean carbon in the ocean interior (Section 5.2.3.1; Gruber et al., 2019; Arora et al., 2020). The land-to-ocean riverine flux and the carbon burial in ocean sediments also play a minor role (*low confidence*) (Arora et al., 2020). Based on *high agreement* of projections by coupled climate models, there is *high confidence* that the resultant climate-carbon cycle feedbacks are positive but the extent of the weakening the ocean sink is scenario dependent (Arora et al., 2020).

Regionally, the Southern Ocean is a major sink of anthropogenic CO<sub>2</sub> (Figure 5.8a), although challenges in modelling its circulation and Antarctic sea ice transport (Section 3.4.1.2; Section 9.2.3.2; Section 9.3.2) generate uncertainty in the response of its sink to future carbon-climate feedbacks. Increased freshwater input may cause a slowdown of the lower overturning circulation, leading to increased Southern Ocean biological carbon storage (Ito et al., 2015); alternatively, increased winds may intensify the overturning circulation, reducing the net CO<sub>2</sub> sink in the Southern Ocean (Bronse laer et al., 2018; Saunders et al., 2018). On centennial timescales, there is thus *low confidence* in the overall effect of intensifying winds in the Southern Ocean on CO<sub>2</sub> uptake.

##### 5.4.4.2 Biological Drivers of Future Ocean Carbon Uptake

While physical drivers control the present-day anthropogenic carbon sink, biological processes are responsible for the majority of the vertical gradient in DIC (natural carbon storage). A small fraction of the organic carbon fixed by primary production (PP) reaches the sea floor, where it can be stored in sediments on geological timescales, making the biological carbon pump (BCP) an important mechanism for very long-term CO<sub>2</sub> storage. Projected reductions in ocean ventilation (Section 9.2.1.4) would lengthen residence time and lead to DIC accumulating in the deep ocean due to organic carbon remineralisation.

1  
2 Since AR5 (Section 6.3.2.5.6), progress has been made in understanding the biological drivers of ocean  
3 carbon uptake in both coupled climate models and observations (SROCC Section 5.2.2.6). Here we focus on  
4 potential feedbacks between biological processes and climate. In CMIP5 models, the direction of modelled  
5 PP in response to increased atmospheric CO<sub>2</sub> concentration and climate warming was unclear (Taucher and  
6 Oschlies, 2011; Laufkoetter et al., 2015). This remains the case in the CMIP6 models; indeed, inter-model  
7 uncertainty has increased in CMIP6 models, compared to CMIP5. The projected global multi-model mean  
8 change in PP in 13 models run under the SSP5–8.5 scenario project is  $-3 \pm 9\%$  (2080–2099 mean values  
9 relative to 1870–1899  $\pm$  the inter-model standard deviation; Kwiatkowski et al., 2020). Under the low-  
10 emission, high-mitigation scenario SSP1–2.6, the global change in PP is  $-0.56 \pm 4\%$ . Observations in the  
11 contemporary period provide little direct constraint on the modelled responses of PP to climate change,  
12 partly due to insufficiently long records (Henson et al., 2016). However, there is some indication of an  
13 emergent constraint on changes in tropical PP based on interannual variability derived from remote sensing  
14 (Kwiatkowski et al., 2017b; Section 5.4.6).

15  
16 In CMIP5 models run under RCP8.5, particulate organic carbon (POC) export flux is projected to decline by  
17 1–12% by 2100 (Taucher and Oschlies 2011; Laufkoetter et al. 2015). Similar values are predicted in 18  
18 CMIP6 models, with declines of 2.5–21.5% (median –14%) or 0.2–2 GtC (median –0.8 GtC) between 1900  
19 and 2100 under the SSP5–8.5 scenario. The mechanisms driving these changes vary widely between models  
20 due to differences in parameterisation of particle formation, remineralisation and plankton community  
21 structure.

22  
23 Ocean warming reduces the vertical supply of nutrients to the upper ocean due to increasing stratification  
24 (Section 9.2.1.4) but may also act to alleviate seasonal light limitation. The projected effect is to decrease PP  
25 at low latitudes and increase PP at high latitudes (Kwiatkowski et al., 2020). Future changes to dust  
26 deposition due to desertification (Mahowald et al., 2017), alterations to the nitrogen cycle (Section 5.3.3.2;  
27 SROCC Section 5.2.3.1.2), and reducing sea ice cover (Ardyna and Arrigo, 2020) all have the potential to  
28 alter PP regionally. Higher ocean temperatures tend to result in higher metabolic rates, although respiration  
29 may increase more rapidly than PP (Boscolo-Galazzo et al., 2018; Brewer, 2019; Cavan et al., 2019). Ocean  
30 warming and reduced PP are expected to result in lower zooplankton abundance, and the expansion of OMZs  
31 may reduce the ability of zooplankton to remineralise POC, thus increasing the efficiency of the BCP and  
32 forming a negative climate feedback (Cavan et al., 2017). Increased microbial respiration due to warming  
33 may result in greater quantities of organic carbon transferred into the dissolved organic carbon pool (Jiao et  
34 al., 2014; Legendre et al., 2015; Roshan and DeVries, 2017) which, while increasing the residence time of  
35 carbon in the ocean, would ultimately reduce the sedimentary burial and hence sequestration on geologic  
36 time scales (Olivarez Lyle and Lyle, 2006).

37  
38 Most models project that smaller phytoplankton are favoured in future ocean conditions (*medium confidence*;  
39 Cabré et al., 2015; Fu et al., 2016; Flombaum et al., 2020) driven by warming water and/or changing nutrient  
40 availability, which would alter the magnitude and efficiency of the BCP by altering the sinking speed,  
41 respiration rate and aggregation fragmentation of sinking particles. There is *low confidence* in the sign of the  
42 resulting feedback: regions in which small phytoplankton dominate may have a more efficient pump,  
43 although the total amount of organic carbon reaching the sea floor is lower (Herndl and Reinthaler, 2013;  
44 Bach et al., 2016; Richardson, 2019). Alternatively, an increase in small phytoplankton could result in a less  
45 efficient pump, due either to a greater fraction of PP being processed through the upper ocean microbial loop  
46 (Jiao et al., 2014) or generation of slower sinking particles (Guidi et al., 2009; Leung et al., 2021). Variable  
47 phytoplankton stoichiometry is predicted to increase the amount of carbon stored via the BCP relative to the  
48 amount of PP, so that fixed stoichiometry models (as in CMIP5) may underestimate cumulative ocean  
49 carbon uptake to 2100 by 0.5–3.5% (2–15 PgC) (RCP8.5 scenario; Kwiatkowski et al., 2018). Other climate  
50 effects such as deoxygenation or ocean acidification could also result in alterations to the magnitude and  
51 efficiency of the BCP (Krumhardt et al., 2019; Raven et al., 2021; Taucher et al., 2021).

52  
53 Based on *high agreement* across multiple lines of evidence and physical understanding there is thus *high*  
54 *confidence* that feedbacks to climate will arise from alterations to the magnitude and efficiency of the BCP  
55 changing PP, and the depth of remineralisation. However, the complexity of the mechanisms involved in the

1 export and remineralisation of POC result in *low confidence* in the magnitude and sign of biological  
 2 feedbacks to climate. Nevertheless, improved model representation of PP and the BCP is required (which  
 3 requires better observational constraints), as the contribution of biological processes to CO<sub>2</sub> uptake is  
 4 expected to become more significant with continued climate change (Hauck et al., 2015).  
 5  
 6

#### 7 5.4.5 Carbon Cycle Projections in Earth System Models

8  
 9 This section summarises future projections of land and ocean carbon sinks from the latest ESMs. ESMs are  
 10 the basis for century timescale projections (Chapter 4), and for detection and attribution studies (Chapter 3).  
 11 These models aim to simulate the evolution of the carbon sources and sinks on land and in the ocean, in  
 12 addition to the physical components of the climate system. ESMs include interactions between many of the  
 13 processes and feedbacks described in Sections 5.4.1 to 5.4.4.  
 14

15 ESMs are now integral to the coupled model intercomparison project. Model output data from CMIP5 was  
 16 analysed in the AR5, while data from CMIP6 forms the basis for the analysis presented in this subsection.  
 17 The CMIP5 ESMs discussed in AR5 (WGI, Section 6.4.2) produced a wide range of projections of future  
 18 CO<sub>2</sub> (Friedlingstein et al., 2014) primarily associated with different magnitudes of carbon-climate and  
 19 carbon-concentration feedbacks (Arora et al., 2013), but also exacerbated by differences in the simulation of  
 20 the net carbon release from land-use change (Brovkin et al., 2013). A key deficiency of almost all CMIP5  
 21 ESMs was the neglect of nutrient limitations on CO<sub>2</sub>-fertilisation of land plant photosynthesis (Zaehle et al.,  
 22 2015) (see Section 5.4.1).  
 23

24 Some CMIP6 models considered in this report now include nitrogen limitations on land vegetation growth,  
 25 along with many other added processes compared to CMIP5. Table 5.4 summarises characteristics of the  
 26 land and ocean carbon cycle models used in CMIP6 ESMs (Arora et al., 2020). In CMIP6, most ocean  
 27 carbon cycle models (8 of 11) track three or more limiting nutrients (most often nitrogen, phosphorus,  
 28 silicon, iron), and represent two or more phytoplankton types. More than half of the land carbon cycle  
 29 models (6 of 11) now include an interactive nitrogen cycle, and almost half (5 of 11) represent forest fires.  
 30 However, even for CMIP6, very few models explicitly represent vegetation dynamics (3 of 11) or permafrost  
 31 carbon (2 of 11). Despite these remaining limitations, the carbon cycle components of CMIP6 represent an  
 32 advance on those in CMIP5 as they represent additional important processes (e.g. nitrogen-limitations on the  
 33 land carbon sink, iron-limitations on ocean ecosystems).  
 34  
 35

36 [START TABLE 5.4 HERE]

37  
 38 **Table 5.4: Properties of the CMIP6 Earth system models (ESMs), focussing on the land and ocean carbon**  
 39 **cycle components of these models** (Arora et al., 2020). Characteristics listed under each ESM are:  
 40 number of vegetation carbon pools (Veg C pools), number of soil and litter carbon pools (Dead C pools),  
 41 number of Plant Functional Types (PFTs), whether forest fire is represented (Fire), whether vegetation  
 42 dynamics is represented (Dynamic Veg), whether permafrost carbon is represented (Permafrost C),  
 43 whether the nitrogen cycle is represented (Nitrogen cycle), the number of phytoplankton types  
 44 (Phytoplankton), the number of zooplankton types (Zooplankton), and the list of ocean nutrients  
 45 represented (Nutrients).  
 46

Modelling group	CSIRO	BCC	CCCma	CESM	CNRM	GFDL	IPSL	JAMSETC	MPI	NorESM2-LM	UK
ESM	ACCESS-ESM1.5	BCC-CSM2-MR	CanESM5	CESM2	CNRM-ESM2-1	GFDL-ESM4	IPSL-CM6A-LR	MIROC-ES2L	MPI-ESM1.2-LR	NorESM2-LM	UKESM1-0-LL
<b>Land carbon/biogeochemistry component</b>											
Model name	CABLE2.4 CASA-CNP	BCC-AVIM2	CLASS-CTEM	CLM5	ISBA-CTRIP	LM4p1	ORCHIDEE (2)	MATSIRO (phys) VISIT-e (BGC)	JSBACH3.2	CLM5	JULES-ES-1.0
Veg C pools	3	3	3	22	6	6	8	3	3	3	3
Dead C pools	6	8	2	7	7	4	3	6	18	7	4
PFTS	13	16	9	22	16	6	15	13	13	21	13

<b>Fire</b>	No	No	No	Yes	Yes	Yes	No	No	Yes	Yes	No
<b>Dynamic Veg</b>	No	No	No	No	No	Yes	No	No	Yes	No	Yes
<b>Permafrost C</b>	No	No	No	Yes	No	No	No	No	No	Yes	No
<b>Nitrogen cycle</b>	Yes	No	No	Yes	No	No	No	Yes	Yes	Yes	Yes
<b>Ocean carbon/biogeochemistry component</b>											
<b>Model name</b>	WOMBAT	MOM4_L40	CMOC (biol)	MARBL	PISCESv2-gas	COBALTv2	PISCES-v2	OECO2	HAMOCC6	HAMOCC5.1	MEDUSA-2.1
<b>Phytoplankton</b>	1	0	1	3	2	2	2	2	2	1	2
<b>Zooplankton</b>	1	0	1	1	2	3	2	1	1	1	2
<b>Limiting Nutrients</b>	P, Fe	P	N	N, P, Si, Fe	N, P, Si, Fe	N, P, Si, Fe	N, P, Si, Fe	N, P, Fe	N, P, Si, FE	N, P, Si, Fe	N, Si, Fe

[END TABLE 5.4 HERE]

ESMs can be driven by anthropogenic CO<sub>2</sub> emissions ('emissions-driven' runs), in which case atmospheric CO<sub>2</sub> concentration is a predicted variable; or by prescribed time-varying atmospheric concentrations ('concentration-driven' runs). In concentration-driven runs, simulated land and ocean carbon sinks respond to the prescribed atmospheric CO<sub>2</sub> and resulting changes in climate, but do not feed back through changes in the atmospheric CO<sub>2</sub> concentration. Concentration-driven runs are used to diagnose the carbon emissions consistent with the SSPs and other prescribed concentration scenarios (Section 5.5). In this subsection we specifically analyse results from concentration-driven ESM projections.

#### 5.2.1.1 Evaluation of the Contemporary Carbon Cycle in Concentration-Driven Runs

Models need to be compared to as wide an array of observational benchmarks as possible in order to have confidence in their projections. This is particularly the case for highly-uncertain land carbon cycle feedbacks (Arora et al., 2013; Friedlingstein et al., 2014b). Land models within ESMs can be compared to multiple different datasets of processes such as gross carbon uptake, physical predictions such as leaf area and carbon stocks which influence carbon fluxes and are diagnostic of carbon turnover times, as well as linkages between carbon and water cycles and other aspects of the terrestrial carbon cycle. To provide these multiple orthogonal constraints, a model benchmarking system, the international land model benchmarking (ILAMB), has been developed (Collier et al., 2018).

Figure 5.22 shows an overview set of land (Figure 5.22a) and ocean (Figure 5.22b) benchmarks applied to both the CMIP5 and CMIP6 historical simulations. There is good evidence of an improvement in model performance from CMIP5 (in yellow) to CMIP6 (in green), in both the land and ocean, based on these benchmarks. The mean of the CMIP6 land models outperform or perform equivalently to the mean of the CMIP5 land models on all available metrics.

[START FIGURE 5.22 HERE]

**Figure 5.22: Overview scores for CMIP5 (left hand side of table) and CMIP6 (right hand side of table) Earth system models (ESMs), for multiple benchmarks against different datasets.** (a) Benchmarking of ESM land models, (b) benchmarking of ocean models. Scores are relative to other models within each benchmark row, with positive scores indicating a better agreement with observations. Models included are only those from institutions that participated in both CMIP5 and CMIP6 carbon cycle experiments, in order to trace changes from one ensemble to the next. CMIP5 models are labels in yellow and CMIP6 in green, with the multi-model ensemble means labelled in white. Further details on data sources and processing are available in the chapter data table (Table 5.SM.6).

[END FIGURE 5.22 HERE]

#### 5.4.5.1 Evaluation of Historical Carbon Cycle Simulations in Concentration-Driven Runs

This section evaluates concentration-driven historical simulations of changes in land and ocean cumulative carbon uptake, against observation-based estimates from GCP (Le Quéré et al., 2018a). For each model, common historical land-use changes were prescribed (Jones et al., 2016a).

Figure 5.23 shows global annual mean values from CMIP6 concentration-driven runs for 1850 to 2014. The ocean carbon cycle models reproduce historical carbon uptake well, with the model range for the global ocean carbon sink in 2014 ( $2.3\text{--}2.7\text{ GtC yr}^{-1}$ ) clustering around the central GCP estimate of  $2.6 \pm 0.5\text{ GtC yr}^{-1}$ . Simulated cumulative ocean carbon uptake (1850–2014) ranges from 110 to 166 GtC, with a model mean of  $131 \pm 17\text{ PgC}$  which is lower than the GCP estimate of  $150 \pm 25\text{ GtC}$  (Figure 5.23a). This suggests that CMIP6 models may slightly underestimate historical ocean carbon uptake (Watson et al., 2020).

The land carbon cycle components of historical ESM simulations show a larger range, with simulated cumulative land carbon uptake (1850–2014) spanning the range from  $-47$  to  $+21\text{ GtC}$ , compared to the GCP estimate of  $-12 \pm 50\text{ GtC}$  (Figure 5.22b). This range is due in part to the complications of simulating the difference between carbon uptake by intact ecosystems and the direct release of carbon due to land-use change (Hajima et al., 2020a). There is *high confidence* that the land continues to dominate the overall uncertainty in the projected response of the global carbon cycle to climate change.

[START FIGURE 5.23 HERE]

**Figure 5.23: CMIP6 Earth system model (ESM) concentration-driven historical simulations for 1850 to 2014, compared to observation-based estimates from the global carbon project (GCP).** Panel (a) cumulative ocean carbon uptake from 1850 (PgC); (b) cumulative land carbon uptake from 1850 (PgC). Only models that simulate both land and ocean carbon fluxes are shown here. Further details on data sources and processing are available in the chapter data table (Table 5.SM.6).

[END FIGURE 5.23 HERE]

#### 5.4.5.2 Evaluation of Latitudinal Distribution of Simulated Carbon Sinks

This distinction between the relatively high fidelity with which the ocean carbon sink is simulated, and the much wider range of simulations of the land carbon sink, is also evident in the zonal distribution of the sinks (Figure 5.24). We compare the ESM simulations to estimates from three atmospheric inversion models: CAMS (Chevallier et al., 2005), CT 2017 (Peters et al., 2007) and MIROC-ATM4 (Saeki and Patra, 2017). The ocean carbon cycle components of CMIP6 ESMs are able to simulate the tropical  $\text{CO}_2$  source and mid-latitude  $\text{CO}_2$  sink, with relatively small model spread (Figure 5.24a). The CMIP6 ensemble (red wedge) simulates a larger ocean carbon sink at  $50^\circ\text{N}$  and a weaker sink in the Southern Ocean, than the inversion estimate, but with some evidence of a reduction in these residual errors compared to CMIP5 (blue wedge). The spread in inversion fluxes arises primarily from differences in the atmospheric  $\text{CO}_2$  measurement networks and from transport model uncertainties.

It has been previously noted that AR5 models tended to overestimate land-uptake in the tropics and underestimate uptake in the northern mid-latitudes, compared to inversion estimates. The inclusion of nitrogen limitations on  $\text{CO}_2$ -fertilisation within CMIP6 models was expected to reduce this discrepancy (Anav et al., 2013). There is indeed some evidence that the CMIP6 ensemble (red wedge in Figure 5.24b) captures the northern land carbon sink more clearly than CMIP5 (blue wedge in Figure 5.24b), but there remains a tendency for the ESMs to place more of the global land carbon sink in the tropics than the mid-latitudes, compared to the inversion estimates. Based on a consistent signal across CMIP6 ESMs, there is *medium confidence* that land carbon cycle models continue to underestimate the northern hemisphere land carbon sink, when compared to estimates from atmospheric inversion (Ciais et al., 2019).

1  
2 [START FIGURE 5.24 HERE]

3  
4 **Figure 5.24: Comparison of modelled zonal distribution of contemporary carbon sinks against atmospheric**  
5 **inversion estimates for 2000–2009, (a) ocean carbon uptake; (b) net land uptake.** Latitude runs from  
6 90°S (i.e. –90°N) to 90°N. Positive uptake represents a carbon sink to ocean/land while negative uptake  
7 represents a carbon source. The land uptake is taken as Net Biome Productivity (NBP) and so includes  
8 net land-use change emissions. The bands show the mean  $\pm 1$  standard deviation across the available  
9 inversions (black bands, 3 models), CMIP5 Earth system models (ESMs) (blue bands, 12 models for the  
10 ocean, 12 models for the land), and CMIP6 ESMs (red bands, 11 models for ocean, 10 models for land).  
11 Further details on data sources and processing are available in the chapter data table (Table 5.SM.6).  
12

13 [END FIGURE 5.24 HERE]

#### 14 15 16 5.4.5.3 Coupled Climate-Carbon Cycle Projections

17  
18 Land and ocean carbon uptake are driven primarily by increases in atmospheric CO<sub>2</sub> (Figure 5.25). As a  
19 result, the evolution of land and ocean carbon sinks differs significantly between the SSP scenarios. Under  
20 scenarios which have greater increases in atmospheric CO<sub>2</sub> (such as SSP5–8.5 and SSP3–7.0) the absolute  
21 values of the sinks are larger, but the fraction of implied emissions taken-up by the sinks declines through  
22 the 21st century. By contrast, scenarios that assume CO<sub>2</sub> stabilisation in the 21st century (such as SSP1–2.6  
23 or SSP2–4.5), have smaller absolute sinks but these sinks take-up an increasing fraction of the implied  
24 emissions (Figure 5.25d). These general principles apply to both the ocean and land carbon sinks.  
25

26 The concentration-driven CMIP6 ESMs agree well on the evolution of the global ocean carbon sink through  
27 the 21st century for four SSP scenarios (Figure 5.26b). The 5-year ensemble mean ocean sink declines to  $0.6$   
28  $\pm 0.2$  GtC yr<sup>-1</sup> by 2100 under SSP1–2.6, and peaks around 2080 at  $5.4 \pm 0.4$  GtC yr<sup>-1</sup> under SSP5–8.5.  
29 Cumulative ocean carbon uptake from 1850 is projected to saturate at approximately  $290 \pm 30$  GtC under  
30 SSP1–2.6, and to reach  $520 \pm 40$  GtC by 2100 under SSP5–8.5 (Figure 5.25e).  
31

32 The ensemble mean changes in land and ocean sinks are qualitatively similar, but the land shows much  
33 higher interannual variability in carbon uptake (Figure 5.25c) and also a much larger spread in the model  
34 projections of cumulative land carbon uptake (Figure 5.25f). The 5-year ensemble mean net land carbon sink  
35 is projected to decline to  $0.4 \pm 1.0$  GtC yr<sup>-1</sup> by 2100 under SSP1–2.6, and to reach around  $5.6 \pm 3.7$  GtC yr<sup>-1</sup>  
36 under SSP5–8.5 (Figure 5.25b). Cumulative net land carbon uptake from 1850 is projected to saturate at  
37 approximately  $150 \pm 35$  GtC under SSP1–2.6, and to reach  $310 \pm 130$  GtC by 2100 under SSP5–8.5.  
38 Significant uncertainty remains in the future of the global land carbon sink, but there has been a notable  
39 reduction in the model spread from CMIP5 to CMIP6.  
40  
41

42 [START FIGURE 5.25 HERE]

43  
44 **Figure 5.25: Modelled evolution of the global land and ocean carbon sinks for 1850 to 2100 in concentration-**  
45 **driven CMIP6 Earth system model (ESM) scenario runs** (SSP1–2.6: blue; SSP2–4.5: skyblue; SSP3–  
46 7.0: yellow; SSP5–8.5: red): (a) prescribed atmospheric CO<sub>2</sub> concentrations; (b) 5-year running mean  
47 ocean carbon sink (GtC yr<sup>-1</sup>); (c) 5-year running mean net land carbon sink (GtC yr<sup>-1</sup>); (d) inferred  
48 cumulative sink fraction of emissions from 1850; (e) change in ocean carbon storage from 1850 (GtC); (f)  
49 change in land carbon storage from 1850 (GtC). Thick lines represent the ensemble mean of the listed  
50 ESM runs, and the error bars represents  $\pm$  one standard deviation about that mean. The grey wedges  
51 represent estimates from the global carbon project (GCP), assuming uncertainties in the annual mean  
52 ocean and net land carbon sinks of  $0.5$  GtC yr<sup>-1</sup> and  $1$  GtC yr<sup>-1</sup> respectively, and uncertainties in the  
53 changes in carbon stores (ocean, land and cumulative total emissions) of  $25$  GtC. The net land carbon  
54 sink is taken as net biome productivity (NBP) and so includes any modelled net land-use change  
55 emissions. Further details on data sources and processing are available in the chapter data table (Table  
56 5.SM.6).  
57

1 [END FIGURE 5.25 HERE]  
 2  
 3

4 Geographical patterns of carbon changes for four SSP scenarios are shown in Figure 5.26, with cleared areas  
 5 (no diagonal lines) showing agreement on the sign of the change by at least 80% of the models. In all  
 6 scenarios the ocean sink is strongest in the Southern Ocean and North Atlantic. The land carbon sink occurs  
 7 primarily where there are present-day forests. In the mid- and high northern latitudes, a carbon sink is  
 8 projected as a result of the combined impacts of increasing CO<sub>2</sub> and warming (see Section 5.4.5.5). Changes  
 9 in land carbon storage in the tropics also depend strongly on the assumed rate of deforestation which varies  
 10 in magnitude across the SSPs, from relatively low rates in SSP1–2.6 to relatively high rates in SSP3–7.0.

11  
 12  
 13 [START FIGURE 5.26 HERE]  
 14

15 **Figure 5.26: Maps of net carbon changes under four Shared Socioeconomic Pathway (SSP) scenarios, as**  
 16 **evaluated from nine CMIP6 Earth system models.** Uncertainty is represented using the simple  
 17 approach (see Cross-Chapter Box Atlas.1 for more information): No overlay indicates regions with high  
 18 model agreement, where ≥80% of models agree with the ensemble mean on the sign of change; diagonal  
 19 lines indicate regions with low model agreement, where <80% of models agree with the ensemble mean  
 20 on the sign of change. On land, this is calculated as the time integral of NBP, for the ocean it is the time-  
 21 integral of air-sea CO<sub>2</sub> gas flux anomalies relative to the pre-industrial. Further details on data sources and  
 22 processing are available in the chapter data table (Table 5.SM.6).  
 23

24 [END FIGURE 5.26 HERE]  
 25  
 26

27 In summary, oceanic and terrestrial carbon sinks are projected to continue to grow with increasing  
 28 atmospheric concentrations of CO<sub>2</sub>, but the fraction of emissions that is taken up by land and ocean is  
 29 expected to decline as the CO<sub>2</sub> concentration increases (*high confidence*). In the ensemble mean, ESMs  
 30 suggest approximately equal global land and ocean carbon uptake for each of the SSP scenarios. However,  
 31 the range of model projections is much larger for the land carbon sink. Despite the wide range of model  
 32 responses, uncertainty in atmospheric CO<sub>2</sub> by 2100 is dominated by future anthropogenic emissions rather  
 33 than carbon-climate feedbacks (*high confidence*).  
 34

#### 35 36 5.4.5.4 Linear Feedback Analysis 37

38 In order to diagnose the causes of the varying time-evolution of carbon sinks, the traditional linear feedback  
 39 approach is adopted (Friedlingstein et al., 2003), as used previously to analyse C<sup>4</sup>MIP (Friedlingstein et al.,  
 40 2006) and CMIP5 models (Arora et al., 2013). Changes in land carbon storage ( $\Delta C_L$ ) and changes in ocean  
 41 carbon storage ( $\Delta C_o$ ) are decomposed into contributions arising from warming ( $\Delta T$ ) and increases in CO<sub>2</sub>  
 42 ( $\Delta CO_2$ ):

$$43 \Delta C_L = \beta_L \Delta CO_2 + \gamma_L \Delta T$$

$$44 \Delta C_o = \beta_o \Delta CO_2 + \gamma_o \Delta T$$

45 where  $\beta_L$  ( $\beta_o$ ) and  $\gamma_L$  ( $\gamma_o$ ) are coefficients that represent the sensitivity of land (ocean) carbon storage to  
 46 changes in CO<sub>2</sub> and global mean temperature respectively. This feedback formalism is one of several that  
 47 have been proposed for analysing climate-carbon cycle feedbacks (Lade et al., 2018).  
 48

49  
 50  
 51 This quasi-equilibrium framework is scenario dependent because of the timescales associated with land and  
 52 ocean carbon uptake, as discussed in AR5 (WGI, Box 6.4). However, it is retained here for traceability with  
 53 AR5. This approach has been used to define a number of emergent constraints on carbon cycle feedbacks  
 54 (Section 5.4.6) and to reconstruct the transient climate response to emissions (TCRE) (Jones and  
 55 Friedlingstein, 2020), as in Section 5.5. In order to minimise the confounding effect of the scenario



dependence,  $\beta$  and  $\gamma$  values are diagnosed from idealised runs in which a 1% per year increase in atmospheric CO<sub>2</sub> concentration is prescribed, as for AR5 (WGI, Box 6.4) (Arora et al., 2013). Values of  $\beta$  are calculated from ‘biogeochemical’ runs in which the prescribed CO<sub>2</sub> increases do not affect climate, and these are then used to isolate  $\gamma$  values in fully-coupled runs in which both climate and CO<sub>2</sub> change (Friedlingstein et al., 2003).

Table 5.5 shows the global land and global ocean values of  $\beta$  and  $\gamma$  for each of the CMIP6 ESMs (Arora et al., 2020). Also shown in the last two rows are the ensemble mean and standard deviation across the ensemble, for CMIP6 and CMIP5. In both ensembles, the largest uncertainties are in the sensitivity of land carbon storage to CO<sub>2</sub> ( $\beta_L$ ) and the sensitivity of land carbon storage to temperature ( $\gamma_L$ ). The more widespread modelling of nitrogen limitations in CMIP6 was expected to lead to reductions in both of these feedback parameters. There is indeed some evidence for that with ensemble mean  $\gamma_L$  moving from  $-58 \pm 38$  GtC K<sup>-1</sup> to  $-33 \pm 33$  GtC K<sup>-1</sup>. Between CMIP5 and CMIP6, there are also reductions in ensemble mean  $\beta_o$  (0.82 to 0.77 GtC ppm<sup>-1</sup>),  $\beta_L$  (0.93 to 0.89 GtC K<sup>-1</sup>) and  $\gamma_o$  ( $-17.3$  to  $-16.9$  GtC K<sup>-1</sup>), but these are progressively less significant compared to the model spread in each case.

[START TABLE 5.5 HERE]

**Table 5.5: Diagnosed global feedback parameters for CMIP6 ESMs based on 4 x CO<sub>2</sub> runs** (Arora et al., 2020). The last two rows show the mean and standard deviation cross the CMIP6 and CMIP5 models, respectively.

Model Name	Land Feedback Factors			Ocean Feedback Factors	
	$\beta_L$ (PgC ppm <sup>-1</sup> )	$\gamma_L$ (PgC K <sup>-1</sup> )		$\beta_o$ (PgC ppm <sup>-1</sup> )	$\gamma_o$ (PgC K <sup>-1</sup> )
ACCESS-ESM1.5	0.37	-21.1		0.90	-23.8
CanESM5	1.28	16.0		0.77	-14.7
CESM2	0.90	-21.6		0.71	-10.9
CNRM-ESM2-1	1.36	-83.1		0.70	-9.4
IPSL-CM6A-LR	0.62	-8.7		0.76	-13.0
MIROC-ES2L	1.12	-69.6		0.73	-22.3
MPI-ESM1.2-LR	0.71	-5.2		0.77	-20.1
NOAA-GFDL-ESM4	0.93	-80.1		0.84	-21.7
NorESM2-LM	0.85	-21.0		0.78	-19.6
UKESM1-0-LL	0.75	-38.4		0.75	-14.1
<b>CMIP6 Model Mean</b>	<b>0.89</b> ± <b>0.30</b>	<b>-33.3</b> ± <b>33.8</b>		<b>0.77</b> ± <b>0.06</b>	<b>-16.9</b> ± <b>5.1</b>
<b>CMIP5 Model Mean</b>	<b>0.93</b> ± <b>0.49</b>	<b>-57.9</b> ± <b>38.2</b>		<b>0.82</b> ± <b>0.07</b>	<b>-17.3</b> ± <b>3.8</b>

[END TABLE 5.5 HERE]

In these idealised 1% per year CO<sub>2</sub> runs, the CMIP6 models show reasonable agreement on the patterns of carbon uptake and also on the separate impacts of CO<sub>2</sub> increase and climate change (Figure 5.27). For the ensemble mean, increasing atmospheric CO<sub>2</sub> increases carbon uptake by the oceans, especially in the

Southern Ocean and the North Atlantic Ocean; and on the land, especially in tropical and boreal forests ( $\beta$ , Figure 5.27a). Climate change further enhances land carbon storage in the boreal zone but has a compensating negative impact on the carbon sink in tropical and subtropical lands, and in the North Atlantic

1 Ocean ( $\gamma$ , Figure 5.27b). Overall, the ensemble mean of the CMIP6 ESMs models indicates increasing  
 2 carbon storage with CO<sub>2</sub> in almost all locations (Figure 5.27c).

3  
 4  
 5 **[START FIGURE 5.27 HERE]**

6  
 7 **Figure 5.27: Maps of carbon-concentration and carbon-climate feedback terms, as well as net carbon changes**  
 8 **under the idealised 1% per year CO<sub>2</sub> scenario, as evaluated from CMIP6 Earth system models**  
 9 **(ESMs).** Shown are the model means from nine CMIP6 ESMs. Uncertainty is represented using the  
 10 simple approach (see Cross-Chapter Box Atlas.1 for more information): No overlay indicates regions  
 11 with high model agreement, where  $\geq 80\%$  of models agree with the ensemble mean on the sign of change;  
 12 diagonal lines indicate regions with low model agreement, where  $< 80\%$  of models agree with the  
 13 ensemble mean on the sign of change. Also shown are zonal-mean latitude profiles of land (green) and  
 14 ocean (blue) feedbacks. On the land, the zonal mean feedback for the mean of the ensemble of models  
 15 that include nitrogen is shown as dashed lines, and carbon-only models as dash-dotted lines, and the  
 16 carbon-climate feedback from one permafrost-carbon enabled ESM is shown as a dotted line carbon  
 17 changes are calculated as the difference between carbon stocks at different times on land and for the  
 18 ocean as the time integral of atmosphere-ocean CO<sub>2</sub> flux anomalies relative to the pre-industrial. The  
 19 denominator for gamma here is the global mean surface air temperature. Further details on data sources  
 20 and processing are available in the chapter data table (Table 5.SM.6).

21  
 22  
 23 **[END FIGURE 5.27 HERE]**

#### 24 25 26 **5.4.6 Emergent constraints to reduce uncertainties in projections**

27  
 28 Emergent constraints are based on relationships between observable aspects of the current or past climate  
 29 (such as trends or variability), and uncertain aspects of future climate change (such as the strength of  
 30 particular feedbacks). These relationships are evident across an ensemble of models. When combined with  
 31 an observational estimate of the trend or variability in the real climate, such emergent relationships can yield  
 32 ‘emergent constraints’ on future climate change (Hall et al., 2019). At the time of the AR5 (WGI, 9.8.3),  
 33 there had been relatively few applications of the technique to constrain carbon cycle sensitivities, but there  
 34 have been many studies published since (see for example summary in Cox, (2019)). Figure 5.28 shows some  
 35 key published emergent constraints on the carbon cycle in ESMs.

36  
 37  
 38 **[START FIGURE 5.28 HERE]**

39  
 40 **Figure 5.28: Examples of emergent constraints on the carbon cycle in Earth system models (ESMs),** reproduced  
 41 from previously published studies: (a) projected global mean atmospheric CO<sub>2</sub> concentration by 2060  
 42 under the RCP8.5 emissions scenario against the simulated CO<sub>2</sub> in 2010 (Friedlingstein et al., 2014b;  
 43 Hoffman et al., 2014); (b) sensitivity of tropical land carbon to warming ( $\gamma_{LT}$ ) against the sensitivity of the  
 44 atmospheric CO<sub>2</sub> growth-rate to tropical temperature variability (Cox et al., 2013b; Wenzel et al., 2014);  
 45 (c) sensitivity of extratropical (30°N–90°N) gross primary production to a doubling of atmospheric  
 46 CO<sub>2</sub> against the sensitivity of the amplitude of the CO<sub>2</sub> seasonal cycle at Kumkahi, Hawaii to global  
 47 atmospheric CO<sub>2</sub> concentration (Wenzel et al., 2016); (d) change in high-latitude (30°N–90°N) gross  
 48 primary production versus trend in high-latitude leaf area index or ‘greenness’ (Winkler et al., 2019); (e)  
 49 sensitivity of the primary production of the Tropical ocean to climate change versus its sensitivity to  
 50 ENSO-driven temperature variability (Kwiatkowski et al., 2017); (f) global ocean carbon sink in the  
 51 2090s versus the current-day carbon sink in the Southern Ocean. In each case, a red-dot represents a  
 52 single ESM projection, the grey bar represents the emergent relationship between the y-variable and the  
 53 x-variable, the blue bar represents the observational estimate of the x-axis variable, and the green bar  
 54 represents the resulting emergent constraint on the y-axis variable. The thicknesses represent  $\pm$  one  
 55 standard error in each case. Figure after Cox, (2019). Further details on data sources and processing are  
 56 available in the chapter data table (Table 5.SM.6).

1 [END FIGURE 5.28 HERE]  
2  
3

#### 4 5.4.7 Climate Feedbacks from CH<sub>4</sub> and N<sub>2</sub>O 5

6 Sources and sinks of CH<sub>4</sub> and N<sub>2</sub>O respond both directly and indirectly to atmospheric CO<sub>2</sub> concentration  
7 and climate change, and thereby give rise to additional biogeochemical feedbacks in the climate system,  
8 which may amplify or attenuate climate-carbon cycle feedbacks (Gasser et al., 2017; Lade et al., 2018;  
9 Denisov et al., 2019). Many of these of feedbacks are only partially understood, and thus were only partially  
10 addressed in AR5 (WGI, Sections 6.3.3, 6.3.4, 6.4.7). Since AR5, a growing body of estimates from ESMs,  
11 as well as independent modelling and observation-based studies, enable improved estimates of the associated  
12 feedbacks.  
13

14 The goal of this section is to assess the climate feedback parameters  $\alpha$ , as it is defined in Section 7.4.1.1, for  
15 CH<sub>4</sub> and N<sub>2</sub>O biogeochemical feedbacks. The strength of the feedbacks is estimated in a linear framework  
16 (Gregory et al., 2009), using the radiative forcing equations for CO<sub>2</sub>, CH<sub>4</sub> and N<sub>2</sub>O (Etminan et al., 2016). In  
17 addition to estimates from ESMs, the feedback parameter  $\alpha$  is estimated from independent estimates of  
18 surface emission climate sensitivities and atmospheric box models, following (Arneeth et al., 2010; Thornhill  
19 et al., 2020). These assessed feedback parameters are used in Section 7.4.2.5.  
20

21 CH<sub>4</sub> feedbacks may arise from changing wetland emissions (including rice farming) and from sources that  
22 are expected to grow under climate change (e.g. related to permafrost thaw, fires, and freshwater bodies).  
23 CH<sub>4</sub> emissions from wetlands and landfills generally increase with warming due to enhanced decomposition  
24 with higher temperatures, thereby potentially providing a positive CH<sub>4</sub> feedback on climate (Dean et al.,  
25 2018). The contribution of wetlands to interannual variability of atmospheric CH<sub>4</sub> is shaped by the different  
26 impacts of temperature and precipitation anomalies on wetland emissions (e.g. during El Niño episodes) and  
27 therefore the relationship between climate anomalies and the wetland contribution to the CH<sub>4</sub> growth rate is  
28 complex (Pison et al., 2013; Nisbet et al., 2016; Zhang et al., 2020b). As assessed by SROCC (IPCC,  
29 2019b), there is *high agreement* across model simulations that wetlands CH<sub>4</sub> emissions will increase in the  
30 21st century, but *low agreement* in the magnitude of the change (Denisov et al., 2013; Shindell et al., 2013a;  
31 Stocker et al., 2013a; Zhang et al., 2017; Koffi et al., 2020). Climate change increases wetland emissions  
32 (Gedney, 2004; Volodin, 2008; Ringeval et al., 2011; Denisov et al., 2013; Shindell et al., 2013a; Gedney et  
33 al., 2019) and gives rise to an estimated wetland CH<sub>4</sub>-climate feedback of  $0.03 \pm 0.01 \text{ W m}^{-2} \text{ }^{\circ}\text{C}^{-1}$  (mean  $\pm$  1  
34 standard deviation; *limited evidence, high agreement*) (Arneeth et al., 2010; Shindell et al., 2013b; Stocker et  
35 al., 2013a; Zhang et al., 2017). The effect of rising CO<sub>2</sub> on productivity, and therefore on the substrate for  
36 methanogenesis, can further increase the projected increase in wetland CH<sub>4</sub> emissions (Ringeval et al., 2011;  
37 Melton et al., 2013). Model projections accounting for the combined effects of CO<sub>2</sub> and climate change  
38 suggest a potentially larger climate feedback ( $0.01\text{--}0.16 \text{ W m}^{-2} \text{ }^{\circ}\text{C}^{-1}$ ; *limited evidence, limited agreement*)  
39 (Gedney et al., 2019; Thornhill et al., 2020). Methane release from wetlands depends on the nutrient  
40 availability for methanogenic and methanotrophic microorganisms that can further modify this feedback  
41 (Stepanenko et al., 2016; Donis et al., 2017; Beaulieu et al., 2019). Methane emissions from thermokarst  
42 ponds and wetlands resulting from permafrost thaw, is estimated to contribute an additional CH<sub>4</sub>-climate  
43 feedback of  $0.01 [0.003\text{--}0.04, 5\text{--}95\text{th percentile range}] \text{ W m}^{-2} \text{ }^{\circ}\text{C}^{-1}$  (*limited evidence, moderate agreement*).  
44

45 Methane release from wildfires may increase by a up to a factor of 1.5 during the 21st century (Eliseev et al.,  
46 2014a, 2014b; Kloster and Lasslop, 2017). However, given the contemporary estimate for CH<sub>4</sub> from  
47 wildfires of no more than  $16 \text{ TgCH}_4 \text{ yr}^{-1}$  (van der Werf et al., 2017; Saunio et al., 2020), this feedback is  
48 small, adding no more than 40 ppb to the atmospheric CH<sub>4</sub> by the end of the 21st century (*medium*  
49 *confidence*). Methane emissions from pan-Arctic freshwater bodies is also estimated to increase by  $16$   
50  $\text{ TgCH}_4 \text{ yr}^{-1}$  in the 21st century (Tan and Zhuang, 2015). Emissions from subsea and permafrost methane  
51 hydrates are not expected to change substantially in the 21st century (Section 5.4.8).  
52

53 Land biosphere models show *high agreement* that long-term warming will increase N<sub>2</sub>O release from  
54 terrestrial ecosystems (Xu-Ri et al., 2012; Stocker et al., 2013b; Zaehle, 2013a; Tian et al., 2019). A positive  
55 land N<sub>2</sub>O climate feedback is consistent with paleo-evidence based on reconstructed and modelled emissions

during the last deglacial period (Schilt et al., 2014; Fischer et al., 2019b; Joos et al., 2020). The response of terrestrial N<sub>2</sub>O emissions to atmospheric CO<sub>2</sub> increase and associated warming is dependent on nitrogen availability (van Groenigen et al., 2011; Butterbach-Bahl et al., 2013; Tian et al., 2019). Model-based estimates do not account for the potentially strong emission increases in Boreal and Arctic ecosystems associated with future warming and permafrost thaw (Elberling et al., 2010; Voigt et al., 2017). There is *medium confidence* that the land N<sub>2</sub>O climate feedback is positive, but *low confidence* in the magnitude (0.02 ± 0.01 W m<sup>-2</sup> °C<sup>-1</sup>).

Climate change will also affect N<sub>2</sub>O production in the ocean (Codispoti, 2010; Freing et al., 2012; Bopp et al., 2013; Rees et al., 2016; Breider et al., 2019). Model projections in the 21st century show a 4–12% decrease in ocean N<sub>2</sub>O emissions under RCP8.5 due to a combination of factors including increased ocean stratification, decreases in ocean productivity, and the impact of increasing atmospheric N<sub>2</sub>O abundance on the air-sea flux, corresponding to an ocean N<sub>2</sub>O climate feedback of  $-0.008 \pm 0.002 \text{ W m}^{-2} \text{ °C}^{-1}$  (*high agreement, limited evidence*) (Martinez-Rey et al., 2015; Landolfi et al., 2017; Battaglia and Joos, 2018). On millennial timescales, the ocean N<sub>2</sub>O climate feedback may be positive, owing to ocean deoxygenation and long-term increases in remineralisation (Battaglia and Joos, 2018b).

Based-on these studies, there is medium confidence that the combined climate feedback parameter for CH<sub>4</sub> and N<sub>2</sub>O is positive, but there is low confidence in the magnitude of the estimate (0.05 (0.02–0.09) W m<sup>-2</sup> °C<sup>-1</sup>, 5–95th percentile range).

#### 5.4.8 Combined Biogeochemical Climate Feedback

This section assesses the magnitude of the combined biogeochemical feedback in the climate system (Figure 5.29) by integrating evidence from: carbon-cycle projections represented in Earth system models (Section 5.4.5.5), independent estimates of CO<sub>2</sub> emissions due to permafrost thaw (Box 5.1) and fire (Section 5.4.3.2), natural CH<sub>4</sub> and N<sub>2</sub>O emissions (Section 5.4.7), and aerosol and atmospheric chemistry (Section 6.3.6). We derive a physical climate feedback parameter  $\alpha$ , as defined in Section 7.4.1.1, for CO<sub>2</sub>-based feedbacks using the linear framework proposed by Gregory et al. (2009), using the radiative forcing equations for CO<sub>2</sub> (Etminan et al., 2016).

[START FIGURE 5.29 HERE]

**Figure 5.29: Estimates of the biogeochemical climate feedback parameter ( $\alpha$ ).** The parameter  $\alpha$  (W m<sup>-2</sup> °C<sup>-1</sup>) for a feedback variable  $x$  is defined as  $\alpha_x = \frac{\partial N}{\partial x} \frac{dx}{dT}$  where  $\frac{\partial N}{\partial x}$  is the change in TOA energy balance in response to a change in  $x$  induced by a change in surface temperature ( $T$ ), as in Section 7.4.1.1. (a) Synthesis of biogeochemical feedbacks from panels (b) and (c). Red (blue) bars correspond to positive (negative) feedbacks increasing (decreasing) radiative forcing at the top of the atmosphere. Bars denote the mean and the error bar represents the 5–95<sup>th</sup> percentile range of the estimates; (b) carbon-cycle feedbacks as estimated by coupled carbon-cycle climate models in the CMIP5 (Arora et al., 2013) and CMIP6 (Arora et al., 2020) ensembles, where dots represent single model estimates, and filled (open) circles are those estimates which do (not) include the representation of a terrestrial nitrogen cycle; (c) Estimates of other biogeochemical feedback mechanisms based on various modelling studies. Dots represent single estimates, and coloured bars denote the mean of these estimates with no weighting being made regarding the likelihood of any single estimate, and error bars the 5–95<sup>th</sup> percentile range derived from these estimates. Results in panel (c) have been compiled from (a) Section 5.4.3.2 (Eliseev et al., 2014a; Harrison et al., 2018); (b) Section 5.4.3.3 (Schneider von Deimling et al., 2012; Burke et al., 2013, 2017b, Koven et al., 2015c, 2015b; MacDougall and Knutti, 2016b; Gasser et al., 2018; Kleinen and Brovkin, 2018), where the estimates from Burke et al. have been constrained as assessed in their study (c) Section 5.4.7 (Schneider von Deimling et al., 2012, 2015; Koven et al., 2015c; Turetsky et al., 2020); (d) Section 5.4.7 (Arneeth et al., 2010; Denisov et al., 2013; Shindell et al., 2013a; Stocker et al., 2013a; Zhang et al., 2017); (f) Section 5.4.7 (Xu-Ri et al., 2012; Zachle, 2013; Stocker et al., 2013; Tian et al., 2019); (g) Section 5.4.7 (Martinez-Rey et al., 2015; Landolfi et al., 2017; Battaglia and Joos, 2018). (h) Section 6.3, Table 6.9 mean and the 5–95<sup>th</sup> percentile range of the assessed feedback parameter. Further details on data

sources and processing are available in the chapter data table (Table 5.SM.6).

**[END FIGURE 5.29 HERE]**

The climate feedback parameter for CO<sub>2</sub> ( $-1.13 \pm 0.45 \text{ W m}^{-2} \text{ }^{\circ}\text{C}^{-1}$ , mean and 1 standard-deviation range) is dominated by the contribution of the CO<sub>2</sub>-induced increase of ocean and land carbon storage ( $-1.46 \pm 0.41 \text{ W m}^{-2} \text{ }^{\circ}\text{C}^{-1}$ , corresponding to a  $\beta_{\text{L+O}}$  of  $1.66 \pm 0.31 \text{ PgC ppm}^{-1}$ ), with smaller contributions from the carbon cycle's response to climate ( $0.24 \pm 0.17 \text{ W m}^{-2} \text{ }^{\circ}\text{C}^{-1}$ , corresponding to  $\gamma_{\text{L+O}}$  of  $-50 \pm 34 \text{ PgC }^{\circ}\text{C}^{-1}$ ), and emissions from permafrost thaw ( $0.09 [0.02-0.20] \text{ W m}^{-2} \text{ }^{\circ}\text{C}^{-1}$ , corresponding to  $\gamma$  of  $-18 [3-41] \text{ PgC }^{\circ}\text{C}^{-1}$ , mean and 5–95<sup>th</sup> percentile range) (Figure 5.29a). This estimate does not include an estimate of the fire-related CO<sub>2</sub> feedback (range:  $0.01-0.06 \text{ W m}^{-2} \text{ }^{\circ}\text{C}^{-1}$ ), as only limited evidence was available to inform its assessment. The sum (mean and 5–95<sup>th</sup> percentile range) of feedbacks from natural emissions of CH<sub>4</sub> including permafrost thaw, and N<sub>2</sub>O ( $0.05 [0.02-0.09] \text{ W m}^{-2} \text{ }^{\circ}\text{C}^{-1}$ ), and feedbacks from aerosol and atmospheric chemistry ( $-0.20 [-0.41 \text{ to } 0.01] \text{ W m}^{-2} \text{ }^{\circ}\text{C}^{-1}$ ) leads to an estimate of the non-CO<sub>2</sub> biogeochemical feedback parameter of  $-0.15 [-0.36 \text{ to } 0.06] \text{ W m}^{-2} \text{ }^{\circ}\text{C}^{-1}$ . There is *low confidence* in the estimate of the non-CO<sub>2</sub> biogeochemical feedbacks, due to the large range in the estimates of  $\alpha$  for some individual feedbacks (Figure 5.29c), which can be attributed to the diversity in how models account for these feedbacks, limited process-level understanding, and the existence of known feedbacks for which there is not sufficient evidence to assess the feedback strength.

CO<sub>2</sub> and non-CO<sub>2</sub> biogeochemical feedbacks are an important component of the assessment of TCRE and the remaining carbon budget (Section 5.5). The feedbacks of the carbon cycle of CO<sub>2</sub> and climate are implicitly taken account in the TCRE assessment, because they are represented in the various underlying lines of evidence. Other feedback contributions, such as the non-CO<sub>2</sub> biogeochemical feedback, can be converted into a carbon-equivalent feedback term ( $\gamma$ ; Section 5.4.5.5, 7.6) by reverse application of the linear feedback approximation (Gregory et al., 2009). The contributions of non-CO<sub>2</sub> biogeochemical feedbacks combine to a linear feedback term of  $30 \pm 27 \text{ PgCeq }^{\circ}\text{C}^{-1}$  (1 standard deviation range,  $111 \pm 98 \text{ Gt CO}_2\text{eq }^{\circ}\text{C}^{-1}$ ), including a feedback term of  $-11 (-18 \text{ to } -5) \text{ PgCeq }^{\circ}\text{C}^{-1}$  (5<sup>th</sup>–95<sup>th</sup> percentile range,  $-40 (-62 \text{ to } -18) \text{ Gt CO}_2\text{eq }^{\circ}\text{C}^{-1}$ ) from natural CH<sub>4</sub> and N<sub>2</sub>O sources. The biogeochemical feedback from permafrost thaw leads to a combined linear feedback term of  $-21 \pm 12 \text{ PgCeq }^{\circ}\text{C}^{-1}$  (1 standard deviation range  $-77 \pm 44 \text{ Gt CO}_2\text{eq }^{\circ}\text{C}^{-1}$ ). For the integration of these feedbacks in the assessment of the remaining carbon budget (Section 5.5.2), two individual non-CO<sub>2</sub> feedbacks (tropospheric ozone, and methane lifetime) are captured in the AR6-calibrated emulators (Box 7.1). Excluding those two contributions, the resulting combined linear feedback term for application in Section 5.5.2 is assessed at a reduction of  $7 \pm 27 \text{ PgCeq }^{\circ}\text{C}^{-1}$  (1 standard deviation range,  $-26 \pm 97 \text{ PgCeq }^{\circ}\text{C}^{-1}$ ). For the same reasons as for the feedback terms expressed in  $\text{W m}^{-2} \text{ }^{\circ}\text{C}^{-1}$  (see above), there is overall *low confidence* in the magnitude of these feedbacks.

#### 5.4.9 Abrupt Changes and Tipping Points

The applicability of the linear feedback framework (Section 5.4.5.5) suggests that large-scale biogeochemical feedbacks are approximately linear in the forcing from changes in CO<sub>2</sub> and climate. Nevertheless, regionally the biosphere is known to be capable of producing abrupt changes or even ‘tipping points’ (Higgins and Scheiter, 2012; Lasslop et al., 2016). Abrupt change is defined as a change in the system that is substantially faster than the typical rate of the changes in its history (Chapter 1, Section 1.4.5). A related matter is a tipping point: a critical threshold beyond which a system reorganizes, often abruptly and/or irreversibly. Possible abrupt changes in the Earth system include those related to ecosystems and biogeochemistry (Lenton et al., 2008; Steffen et al., 2018): tropical and boreal forest dieback; and release of greenhouse gases from permafrost and methane clathrates (Table 5.6). It is not currently possible to carry out a full assessment of proposed abrupt changes and tipping points in the biogeochemical cycles. In this section we therefore focus instead on estimating upper limits on the possible impact of abrupt changes on the evolution of atmospheric GHGs out to 2100, for comparison to the impact of direct anthropogenic emissions.

1 [START TABLE 5.6 HERE]

2  
3 **Table 5.6: Examples of possible biogeochemical abrupt changes and tipping points in the Earth system.** The  
4 fourth and sixth comments provide upper estimates of the impact of each example on the evolution of  
5 atmospheric GHGs in the 21st century. These upper estimates are therefore *very unlikely* but provide a  
6 useful comparison to the impact of direct anthropogenic emissions (currently 2.5 ppm yr<sup>-1</sup>).  
7

Abrupt change/Tipping point	Key region(s)	Probability to occur in the 21st century	Maximum carbon dioxide or methane release in the 21st century	Principal development timescale	Maximum CO <sub>2</sub> or CH <sub>4</sub> rate of change over the 21st century	(Ir)reversibility
Tropical forests dieback (Section 5.4.9.1.1)	Amazon watershed	low	<200 PgC as CO <sub>2</sub> (Section 5.4.8.1.1; <i>medium confidence</i> )	multi-decadal	CO <sub>2</sub> : <0.5 ppm yr <sup>-1</sup>	irreversible at multi-decadal scale ( <i>medium confidence</i> )
Boreal forests dieback (Section 5.4.9.1.1, 5.4.3.2)	boreal Eurasia and North America	low	<27 Pg (Section 5.4.8.1.1; <i>medium confidence</i> )	multi-decadal	small ( <i>low confidence</i> )	irreversible at multi-decadal scale ( <i>medium confidence</i> )
Biogenic emissions from permafrost thaw (Section 5.4.9.1.2)	pan-Arctic	high	up to 240 PgC of CO <sub>2</sub> and up to 5300 Tg of CH <sub>4</sub> (Section 5.4.8.1.2; <i>low confidence</i> )	multi-decadal	CO <sub>2</sub> : ≤1 ppm yr <sup>-1</sup> CH <sub>4</sub> : ≤10 ppb yr <sup>-1</sup>	irreversible at centennial timescales ( <i>high confidence</i> )
Methane release from clathrates (Section 5.4.9.1.3)	oceanic shelf	very low	<i>very likely</i> small (Section 5.4.9.1.3)	multi-millennium	CH <sub>4</sub> : ≤0.2 ppb yr <sup>-1</sup>	irreversible at multi-millennium timescales ( <i>medium confidence</i> )

8 [END TABLE 5.6 HERE]

9  
10  
11  
12 *5.4.9.1 Assessment of biogeochemical tipping points*

13  
14 *5.4.9.1.1 Forest Dieback*

15 Published examples of abrupt biogeochemical changes in models include tropical rain forest dieback (Cox et al., 2004; Jones et al., 2009; Brando et al., 2014; Le Page et al., 2017; Zemp et al., 2017), and temperate and  
16 boreal forest dieback (Joos et al., 2001; Lucht et al., 2006; Scheffer et al., 2012; Lasslop et al., 2016) (see  
17 also Section 5.4.3). Such transitions may be related to (i) large-scale changes in mean climate conditions  
18 crossing particular climate thresholds (Joos et al., 2001; Cox et al., 2004; Lucht et al., 2006; Hirota et al.,  
19 2011; Scheffer et al., 2012; Le Page et al., 2017; Zemp et al., 2017), (ii) temperature and precipitation  
20 extremes (Staver et al., 2011; Higgins and Scheiter, 2012; Scheffer et al., 2012; Pavlov, 2015; Zemp et al.,  
21 2017), or (iii) possible enhancement and intermittency in fire activity (Staver et al., 2011; Higgins and  
22 Scheiter, 2012; Lasslop et al., 2016; Brando et al., 2020). Simulated changes in forest cover are a  
23 combination of the effects of CO<sub>2</sub> on photosynthesis and water-use efficiency (Section 5.4.1), and the effects  
24 of climate change on photosynthesis, respiration and disturbance (Section 5.4.3). In ESMs, direct CO<sub>2</sub>  
25 effects tend to enhance forest growth, but the impacts of climate change vary between being predominantly  
26 negative in the tropics and predominantly positive in the boreal zone (Figure 5.27).  
27  
28

29 Most ESMs project continuing carbon accumulation in tropical forests as a result of direct CO<sub>2</sub> effects  
30 overwhelming the negative effects of climate change (Huntingford et al., 2013; Drijfhout et al., 2015;

1 Boulton et al., 2017). In the real world, forests may be less vulnerable to climate changes than those  
2 modelled in ESMs because of the greater plant trait diversity which confers additional resilience (Reyer et  
3 al., 2015; Levine et al., 2016; Sakschewski et al., 2016), and also because of possible acclimation of  
4 vegetation to warming (Good et al., 2011, 2013; Lloret et al., 2012; Mercado et al., 2018). Contrary, forests  
5 may be more vulnerable in the real world due to indirect climate change effects such as insect outbreaks and  
6 diseases not considered here (Section 5.4.3.2) or model limitations in representing the effects disturbances  
7 such as wildfire and droughts. In general, forests are most vulnerable when climate change is combined with  
8 increased rates of direct deforestation (Nobre et al., 2016; Le Page et al., 2017).

9  
10 In order to estimate an upper limit on the impact of Amazon forest dieback on atmospheric CO<sub>2</sub>, we consider  
11 the *very unlikely* limiting case of negligible direct-CO<sub>2</sub> effects (Section 5.4.1). Emergent constraint  
12 approaches (Section 5.4.6) may be used to estimate an overall loss of tropical land carbon due to climate  
13 change alone, of around 50 PgC per °C of tropical warming (Cox et al., 2013b; Wenzel et al., 2014). This  
14 implies an upper limit to the release of tropical land carbon of <200 PgC over the 21st century (assuming  
15 tropical warming of <4°C, and no CO<sub>2</sub>-fertilisation), which translates to dCO<sub>2</sub>/dt < 0.5 ppm yr<sup>-1</sup>.

16  
17 Boreal forest dieback is not expected to change the atmospheric CO<sub>2</sub> concentration substantially because  
18 forest loss at the south is partly compensated by (i) temperate forest invasion into the previous boreal area  
19 and (ii) boreal forest gain at the north (Friend et al., 2014; Kicklighter et al., 2014; Schaphoff et al., 2016)  
20 (*medium confidence*). An upper estimate of this magnitude, based on statistical modelling of climate change  
21 alone, is of 27 Pg vegetation C loss in the southern boreal forest, which is roughly balanced by gains in the  
22 northern zone (Koven, 2013). Carbon release from vegetation and soil due to wildfires in boreal regions  
23 (Eliseev et al., 2014b; Turetsky et al., 2015; Walker et al., 2019a) is also not expected to change this estimate  
24 substantially because of its small present-day value of about 0.2 PgC yr<sup>-1</sup> (van der Werf et al., 2017), and  
25 because of *likely* increases in precipitation in boreal regions (Chapter 4, Section 4.5.1).

#### 26 27 28 5.4.9.1.2 *Biogenic Emissions Following Permafrost Thaw*

29 There is large uncertainty in release of GHGs from permafrost in the 21st century with the largest of these  
30 estimates implying tens to hundreds of gigatons of carbon released in the form of CO<sub>2</sub> (Box 5.1) and  
31 methane emissions up to 100 TgCH<sub>4</sub> yr<sup>-1</sup> (Box 5.1). A carbon dioxide release of such magnitude would lead  
32 to an increase in the CO<sub>2</sub> accumulation rate in the atmosphere of ≤1 ppm yr<sup>-1</sup>. These emissions develop at  
33 multi-decadal timescale. Assuming a CH<sub>4</sub> lifetime in the atmosphere of the order of 10 years and the  
34 associated feedback parameter of 1.34 ± 0.04 (Chapter 6, Section 6.2.2.1), this would increase the  
35 atmospheric methane content by about 500 ppb over the century, corresponding to a rate of ≤10 ppb yr<sup>-1</sup>.  
36 Irrespective of its origin, additional methane accumulation of such a magnitude is not expected to modify the  
37 temperature response to anthropogenic emissions by more than a few tens of °C cent (Gedney, 2004; Eliseev  
38 et al., 2008; Denisov et al., 2013). Emissions from the permafrost are assessed in Box 5.1.

#### 39 40 41 5.4.9.1.3 *Methane Release from Clathrates*

42 The total global clathrate reservoir is estimated to contain 1500–2000 PgC (Archer et al., 2009; Ruppel and  
43 Kessler, 2017), held predominantly in ocean sediments with only an estimated 20 PgC in and under  
44 permafrost (Ruppel, 2015). The present-day methane release from shelf clathrates is <10 TgCH<sub>4</sub> yr<sup>-1</sup>  
45 (Kretschmer et al., 2015; Saunio et al., 2020). Despite polar amplification (Chapter 7), substantial releases  
46 from the permafrost-embedded subsea clathrates is *very unlikely* (Minshull et al., 2016; Malakhova and  
47 Eliseev, 2017, 2020). This is consistent with an overall small release of methane from the shelf clathrates  
48 during the last deglacial despite large reorganisations in climate state (Bock et al., 2017; Petrenko et al.,  
49 2017; Dyonisius et al., 2020). The long timescales associated with clathrate destabilisation makes it *unlikely*  
50 that CH<sub>4</sub> release from the ocean to the atmosphere will deviate markedly from the present-day value through  
51 the 21st century (Hunter et al., 2013), corresponding to no more than additional 20 ppb of atmospheric  
52 methane (i.e. <0.2 ppb yr<sup>-1</sup>). Another possible source of CH<sub>4</sub> are gas clathrates in deeper terrestrial  
53 permafrost and below it (Buldovicz et al., 2018; Chuvilin et al., 2018), which may have caused recent craters  
54 in the north of Russia (Arzhanov et al., 2016, 2020; Arzhanov and Mokhov, 2017; Kizyakov et al., 2017,  
55 2018). Land clathrates are formed at depths >200 m (Ruppel and Kessler, 2017; Malakhova and Eliseev,

2020), which precludes a substantial response to global warming over the next few centuries and associated emissions.

Thus, it is *very unlikely* that CH<sub>4</sub> emissions from clathrates will substantially warm the climate system over the next few centuries.

#### 5.4.9.2 Abrupt Changes Detected in ESM Projections

Projecting abrupt changes is intrinsically difficult, because by definition abrupt changes occur in a small region of the parameter and/or forcing space. At the time of the AR5 there was no available systematic study of abrupt changes or tipping points in ESMs. An analysis of ESMs since the AR5 has identified a number of abrupt changes in the CMIP5 ensemble (Drijfhout et al., 2015; Bathiany et al., 2020). These include abrupt changes in tropical forests and high-latitude greening, permafrost thaw, and vegetation composition change (Bathiany et al., 2020). Most modelled abrupt changes were detected in boreal and tundra regions, with few models showing Amazon forest dieback (Bathiany et al., 2020).

Based on the evidence presented in this section, we conclude that abrupt changes and tipping points in the biogeochemical cycles lead to additional uncertainty in 21st century GHG concentrations changes, but these are *very likely* to be small compared to the uncertainty associated with future anthropogenic emissions (*high confidence*).

#### 5.4.10 Long Term Response past 2100

AR5 assessed with *very high confidence* that the carbon cycle in the ocean and on land will continue to respond to climate change and rising atmospheric CO<sub>2</sub> concentrations created during the 21st century (WGI, Chapter 6, Executive Summary). Since AR5, experiments with the CESM1 model under the RCP8.5 extension scenario out to 2300, suggest that both land and ocean carbon-climate feedbacks strengthen in time, land and ocean carbon-concentration feedbacks weaken, and the relative importance of ocean sinks versus land sinks increases (Randerson et al., 2015). Under high emissions scenarios, this relative strengthening of land carbon-climate feedbacks leads the terrestrial biosphere to shift from sink to source at some point after 2100 in all of the CMIP5 ESMs and CMIP5-era EMICs (Tokarska et al., 2016). The strengthening of land and ocean carbon-climate feedbacks projected beyond 2100 under high emissions scenarios offsets the declining climate sensitivity to incremental increases of CO<sub>2</sub>, leading to a net strengthening of carbon cycle feedbacks, as measured by the gain parameter, from one century to the next (Randerson et al., 2015).

Figure 5.30 shows carbon cycle changes to 2300 under three SSP scenarios with long-term extensions: SSP5–8.5, SSP5–3.4–overshoot, and SSP1–2.6, for four CMIP6 ESMs and one EMIC. Under all three scenarios, all five models project a reversal of the terrestrial carbon cycle from a sink to a source. However, the reasons for these reversals under very high emissions and low/negative emissions are very different. Under the SSP5–8.5 scenario, the terrestrial carbon-climate feedback is projected to strengthen, while the carbon-concentration feedbacks weaken after emissions peak at 2100, which together drives the land to become a net carbon source after 2100 (Tokarska et al., 2016). The difference in both timing and magnitude of this transition across the ensemble, leads to an assessment of *medium confidence* in the likelihood and *low confidence* in the timing and strength, of the land transitioning from a net sink to a net source under such a scenario. Based on *high agreement* across all available models, we assess with *high confidence* that the ocean sink strength would weaken but not reverse under a long-term high-emissions scenario. In the SSP5–3.4–overshoot scenario, both the terrestrial and ocean reservoirs act as transient carbon sources during the overshoot period, when net anthropogenic CO<sub>2</sub> emissions are negative and CO<sub>2</sub> concentrations are falling, and then revert to near-zero (land) or weak sink (ocean) fluxes after stabilisation of atmospheric CO<sub>2</sub>. The SSP1–2.6 scenario, characterised by lower peak CO<sub>2</sub> concentrations, a smaller overshoot, and much less carbon loss from land use change, shows instead a relaxation towards a neutral biosphere on land and a sustained weak sink in the ocean (see also Section 5.6.2.2.1.2).



1  
2  
3 **[START FIGURE 5.30 HERE]**  
4

5 **Figure 5.30: Trajectories of carbon cycle dynamics for models beyond 2100.** Shown are three scenarios, SSP5–8.5,  
6 SSP5–3.4–overshoot, and SSP1–2.6, from four ESMs (CanESM5, UKESM1, CESM2-WACCM, IPSL-  
7 CM6a-LR) and one EMIC (UVIC-ESCM, (Mengis et al., 2020)) for which extensions beyond 2100 are  
8 available. Solid lines represent the median flux value across the ensemble, and shading represents 15th–  
9 85th percentiles across the ensemble. Further details on data sources and processing are available in the  
10 chapter data table (Table 5.SM.6).  
11

12 **[END FIGURE 5.30 HERE]**  
13  
14

#### 15 *5.4.11 Near-Term Prediction of Ocean and Land Carbon Sinks*

16  
17 The IPCC AR5 (WGI, Section 11.3.2) assessed near-term climate predictability based on ESMs initialised  
18 from the observed climate state. Since the AR5, a growing number of prediction systems have been  
19 developed based on ESMs that include the ocean and land carbon cycle components. Predictability of key  
20 physical climate variables (assessed in Chapter 4) provides a platform to establish predictive skill for  
21 interannual variations in the strength of the natural carbon sinks in response to internal climate variability. In  
22 most systems the carbon cycle components are only indirectly initialised and respond to the initialised  
23 climate variations (Li et al., 2019). This subsection synthesises information on predictability of the land and  
24 ocean carbon sinks using both the idealised potential predictability and the actual predictability skill  
25 measures.  
26

27 Longer-term memory residing in the ocean enables predictability of the ocean carbon sink (McKinley et al.,  
28 2017; Li and Ilyina, 2018). The predictive horizon of the globally integrated air-sea CO<sub>2</sub> fluxes has been  
29 assessed in perfect-model frameworks that are based on an idealised ensemble of simulations in which each  
30 ensemble member serves as a verification, while no observations are assessed. Perfect-model studies provide  
31 an estimate of the upper range of potential predictability for the integrated air-sea CO<sub>2</sub> fluxes of about 2  
32 years globally and up to a decade in some regions (Séférian et al., 2018a; Spring and Ilyina, 2020). Evidence  
33 is also emerging for predictive skill of the global air-sea CO<sub>2</sub> fluxes of up to 6 years based-on prediction  
34 systems initialised with observed physical climate states (Li et al., 2019), with a potential for even longer-  
35 term regional predictability in some regions including the North Atlantic and subpolar Southern Ocean (Li et  
36 al., 2016a; Lovenduski et al., 2019b).  
37

38 Models suggest that predictability of the air-sea CO<sub>2</sub> flux is related to predictability of ocean biogeochemical  
39 state variables such as dissolved inorganic carbon (DIC) and total alkalinity (TA) (Lovenduski et al., 2019b),  
40 as well as the mixed layer depth (Li et al., 2016a). Temperature variations largely control shorter-term  
41 predictability of the ocean carbon sink, while longer term predictability is related to non-thermal drivers such  
42 as ocean circulation and biology (Li et al., 2019). Although there is a substantial spatial heterogeneity,  
43 initialised predictions suggest stronger multi-year variations of the air-sea CO<sub>2</sub> flux and generally tend to  
44 outperform uninitialized simulations on the global scale (Li et al., 2019). The predictive skill of air-sea CO<sub>2</sub>  
45 flux shows a consistent spatial pattern in different models despite the wide range of techniques used to  
46 assimilate observational information (Regnier et al., 2013a). ESM-based prediction systems also demonstrate  
47 predictability of other marine biogeochemical properties such as net primary production (Séférian et al.,  
48 2014; Yeager et al., 2018; Park et al., 2019) and seawater pH (Brady et al., 2020).  
49

50 Seasonal predictability of air-land CO<sub>2</sub> flux up to 6–8 months is driven by the state of ENSO (Zeng et al.,  
51 2008; Betts et al., 2018). Fewer land carbon initialised predictions are available from decadal prediction  
52 systems, yet they tend to outperform the uninitialized simulations in capturing the major year-to-year  
53 variations as indicated by higher correlations with the Global Carbon Budget estimates. There is growing  
54 evidence that potential predictive skill of air-land CO<sub>2</sub> flux is maintained out to a lead-time of 2 years  
55 (Lovenduski et al., 2019a); this predictability horizon is also supported by perfect model studies (Séférian et  
56 al., 2018a; Spring and Ilyina, 2020). The origins of this interannual predictability are not yet fully

1 understood. However, they seem to be associated with the oscillatory behaviour of ENSO (Séférian et al.,  
2 2014) and the predictability of terrestrial ecosystem drivers such as ecosystem respiration and gross primary  
3 production (Lovenduski et al., 2019b). Initialised simulations suggest that observed variability in the land  
4 carbon sink is improved through initialisation of prediction systems with the observed state of the physical  
5 climate.

6  
7 The predictability horizon of variations in atmospheric CO<sub>2</sub> growth-rate are not yet fully established in the  
8 literature. However, predictive skill of the land and ocean carbon sinks show a potential to establish  
9 predictability of variations in atmospheric CO<sub>2</sub> up to 2 years in advance in the initialised prediction systems  
10 with an upper bound of up to 3 years in a perfect-model study (Spring and Ilyina, 2020); this skill is  
11 primarily limited by the terrestrial carbon sink predictability.

## 14 5.5 Remaining Carbon Budgets

15  
16 Science at the time of the IPCC AR5 established a near-linear relationship between cumulative emissions of  
17 CO<sub>2</sub> and the resulting global warming (Allen et al., 2009; Matthews et al., 2009; Meinshausen et al., 2009;  
18 Zickfeld et al., 2009; Collins et al., 2013; Stocker et al., 2013;). The amount of global warming per unit of  
19 cumulated CO<sub>2</sub> emissions is called the transient climate response to cumulative emissions of carbon dioxide  
20 (TCRE). This TCRE relationship is now used to estimate the amount of CO<sub>2</sub> emissions that would be  
21 consistent with limiting global warming to specific levels (Allen et al., 2009; Matthews et al., 2009;  
22 Meinshausen et al., 2009; Zickfeld et al., 2009; Matthews et al., 2012; Collins et al., 2013; Stocker et al.,  
23 2013; Knutti and Rogelj, 2015; Rogelj et al., 2016; Goodwin et al., 2018; Rogelj et al., 2019). The remainder  
24 of CO<sub>2</sub> emissions that would be in line with limiting global warming to a specific temperature level (while  
25 accounting for all other factors affecting global warming) can be estimated with help of the TCRE and is  
26 referred to as the remaining carbon budget (Rogelj et al., 2019; Matthews et al., 2020). Section 5.5.1 first  
27 assesses the TCRE as one of the core concepts underlying the notion of a remaining carbon budget and  
28 Section 5.5.2 then integrates this with the assessment of other contributing factors from across this  
29 assessment to provide a consolidated assessment following the approach of the IPCC SR1.5 (Rogelj et al.,  
30 2018b). The historical carbon budget of CO<sub>2</sub> already emitted is assessed in Section 5.2.1.5.

### 33 5.5.1 Transient Climate Response to Cumulative Emissions of carbon dioxide (TCRE)

#### 35 5.5.1.1 Contributing Physical Processes and Theoretical Frameworks

36  
37 The processes that translate emissions of CO<sub>2</sub> into a change in global temperature (terrestrial and oceanic  
38 carbon uptake, radiative forcing from CO<sub>2</sub>, and ocean heat uptake) are governed by complex mechanisms  
39 that all evolve in time (Gregory et al., 2009) (Sections 3.5, 4.3, 4.5, 5.4, and 7.3, Cross-Chapter Box 5.3).  
40 Starting with an initial description in AR5 (Collins et al., 2013a; Stocker et al., 2013c) a body of literature  
41 has since expanded the understanding of physical mechanisms from which a simple proportional relationship  
42 between cumulative emissions of CO<sub>2</sub> and change in global temperature (expressed in either global mean  
43 surface temperature, GMST or global mean surface air temperature, GSAT) arises.

44  
45 Studies have focused on two key features of the TCRE relationship: (i) why the relationship is nearly  
46 constant in time (Goodwin et al., 2015; MacDougall and Friedlingstein, 2015; Williams et al., 2016; Ehlert et  
47 al., 2017; Katavouta et al., 2018); and (ii) why and under which conditions the relationship is independent on  
48 the historical rate (or pathway) of CO<sub>2</sub> emissions (MacDougall, 2017; Seshadri, 2017).

49  
50 There is increased confidence in the near-constancy of TCRE because of the variety of methods that have  
51 been used to examine this relationship: sensitivity studies with Earth system models of EMICs (Herrington  
52 and Zickfeld, 2014; Ehlert et al., 2017); theory-based equations used to examine Earth system model (ESM)  
53 and EMIC output (Goodwin et al., 2015; Williams et al., 2016, 2017c); and simple analytical models that  
54 capture aspects of the TCRE relationship (MacDougall and Friedlingstein, 2015). All studies agree that the  
55 near-constancy of the TCRE arises from compensation between the diminishing sensitivity of radiative

1 forcing to CO<sub>2</sub> at higher atmospheric concentration, and the diminishing ability of the ocean to take up heat  
2 and carbon at higher cumulative emissions (Allen et al., 2009; Matthews et al., 2009; Frölicher and Paynter,  
3 2015; Goodwin et al., 2015; Gregory et al., 2015; MacDougall and Friedlingstein, 2015; MacDougall, 2016;  
4 Tokarska et al., 2016; Ehlert et al., 2017).

5  
6 The question whether and under which conditions the TCRE relationship is independent of the historical rate  
7 of CO<sub>2</sub> emissions (also referred to as ‘pathway independence of TCRE’) has been examined by using simple  
8 mathematically-tractable models (MacDougall, 2017; Seshadri, 2017). Based on the assumption that the  
9 cumulative fraction of carbon taken up by the terrestrial biosphere is constant, and that the climate feedback  
10 parameter and ocean heat uptake efficacy do not change in time, both studies agree that pathway  
11 independence is sensitive to the rate of CO<sub>2</sub> emissions, such that pathway independence is expected to  
12 breakdown at both very high and very low absolute CO<sub>2</sub> emission rates (MacDougall, 2017; Seshadri, 2017).  
13 Note that in pathways with strongly declining emissions, the cumulative sink fraction by the combined  
14 terrestrial biosphere and ocean is expected to increase (see Figure 5.25 in Section 5.4.5). The studies also  
15 agree that no similar relationship analogous to TCRE can be expected for short-lived non-CO<sub>2</sub> forcings, for  
16 which the annual emissions are a closer proxy for the implied warming (Collins et al., 2013a); Sections 6.4,  
17 7.6. MacDougall, (2017) suggests that two additional constraints are required to create pathway  
18 independence: first, the transport of heat and carbon into the deep ocean should be governed by processes  
19 with similar timescales; and second, the ratio of the net change in the atmospheric carbon pool to the net  
20 change in the ocean carbon pool should be close to the ratio of the enhanced longwave radiation to space (i.e.  
21 the radiative response of the surface) to ocean heat uptake. If these ratios are identical then TCRE would be  
22 completely path independent (MacDougall, 2017). If the ratios are close but not identical, TCRE would be  
23 only approximately path independent over a wide range of cumulative emissions (MacDougall, 2017)  
24 (Cross-Chapter Box 5.3).

25  
26 The land carbon cycle does not appear to play a fundamental role in the origin of the linearity and path-  
27 independence of TCRE (Goodwin et al., 2015; MacDougall and Friedlingstein, 2015; Ehlert et al., 2017),  
28 but, in contrast to the ocean sink, dominates the uncertainty in the magnitude of TCRE by modulating the  
29 cumulative airborne fraction of carbon (Goodwin et al., 2015; Williams et al., 2016; Katavouta et al., 2018;  
30 Jones and Friedlingstein, 2020). Some terrestrial carbon cycle feedbacks (such as the permafrost carbon  
31 feedback, Section 5.4.8, Box 5.1) have the potential to alter both the linearity and pathway independence of  
32 TCRE, if such feedbacks significantly contribute carbon to the atmosphere (MacDougall and Friedlingstein,  
33 2015) (Section 5.5.1.2.3, Section 5.4.8, Box 5.1). A recent study also shows how the value of TCRE can  
34 depend on the effect of ocean ventilation modulating ocean heat uptake (Katavouta et al., 2019).

### 37 [START CROSS-CHAPTER BOX 5.3]

#### 39 **Cross-Chapter Box 5.3: The Ocean Carbon-Heat Nexus and Climate Change Commitment**

41 **Contributors:** Pedro M.S. Monteiro (South Africa), Jean-Baptiste Sallée (France), Piers Foster (UK),  
42 Baylor Fox-Kemper (USA), Helen Hewitt (UK), Masao Ishii (Japan), Joeri Rogelj (Belgium), Kirsten  
43 Zickfeld (Canada/Germany)

#### 45 *Context*

46  
47 In the past 60 years the ocean has taken up and stored  $23 \pm 5\%$  of anthropogenic carbon emissions (*medium*  
48 *confidence*; Section 5.2.1.3) as well as more than 90% of the heat that has accumulated in the Earth system  
49 (referred to as excess heat) since the 1970s (Sections 7.2.2, 9.2.2, 9.2.3; Box 7.2;) (Frölicher et al., 2015;  
50 Talley et al., 2016; Gruber et al., 2019a; Hauck et al., 2020). The interplay between heat and CO<sub>2</sub> uptake by  
51 the ocean has not only played a major role in slowing the rate of global warming but provides a first order  
52 influence in determining the unique properties of a metric of the coupled climate-carbon cycle response,  
53 transient climate response to cumulative carbon emissions (TCRE), which is critical to setting the future  
54 remaining carbon emissions budget (Sections 5.5.1.3, 5.5.4). This role of the ocean in the uptake of heat and  
55 anthropogenic CO<sub>2</sub> and related feedbacks is what we term here the “Ocean Heat-Carbon Nexus”. The ocean

1 processes behind this nexus are important in shaping and understanding the near-linear relationship between  
2 cumulative CO<sub>2</sub> emissions and global warming (TCRE) as well as the uncertainties in future projections of  
3 TCRE properties (Zickfeld et al., 2016; Bronselaer and Zanna, 2020; Jones and Friedlingstein, 2020), its path  
4 independence (MacDougall, 2017), and the warming commitment after cessation of greenhouse gas  
5 emissions (ZEC) (Zickfeld et al., 2016; Ehlert and Zickfeld, 2017; Section 5.5.2). In this box, we assess the  
6 role of the ocean and its physical and chemical thermodynamic processes that shape these striking  
7 characteristics.

8  
9 The role of the ocean in setting the coupled climate-carbon cycle response is threefold: firstly, the ocean and  
10 land carbon sinks together set the airborne fraction (AF) of CO<sub>2</sub> in the atmosphere, which sets the radiative  
11 forcing that drives the additional heat in the atmosphere most of which is taken up by the ocean (Katavouta  
12 et al., 2019; Williams et al., 2019; Sections 7.2, 9.2). However, the land carbon sink does not appear to play  
13 an important role in determining the linearity and path-independence of TCRE (Section 5.5.1.1; Goodwin et  
14 al., 2015; MacDougall and Friedlingstein, 2015; Ehlert et al., 2017). Secondly, the ocean sets the thermal  
15 response through ocean heat uptake (Section 9.2; Frölicher et al., 2015; Bronselaer and Zanna, 2020).  
16 Thirdly, there is a feedback within the ocean heat-carbon nexus as ocean warming, particularly under low or  
17 no mitigation scenarios weakens the ocean sink of CO<sub>2</sub>, which influences the AF, and hence the radiative  
18 forcing (Williams et al., 2019) (Box 7.1). The near-linear relationship between cumulative CO<sub>2</sub> emissions  
19 and global warming (TCRE) is thought to arise to a large extent from the compensation between the  
20 decreasing ability of the ocean to take up heat and CO<sub>2</sub> at higher cumulative CO<sub>2</sub> emissions, pointing to  
21 similar processes that determine ocean uptake of heat and carbon (Goodwin et al., 2015; MacDougall and  
22 Friedlingstein, 2015; Williams et al., 2016; Zickfeld et al., 2016; Ehlert et al., 2017; Section 5.5.1.1).

#### 23 *Processes that drive the Ocean Carbon - Heat Nexus and its change*

24  
25  
26 The air-sea flux of heat and all gases across the ocean interface is driven by a common set of complex and  
27 difficult to observe turbulent diffusion and mixing processes (Wanninkhof et al., 2009; Wanninkhof, 2014;  
28 Cronin et al., 2019; Watson et al., 2020; Sections 5.2.1.3; 9.2.1.2). These processes are typically simplified  
29 into widely verified expressions that link the flux to wind stress, the solubility and the gradient across the air-  
30 sea interface (*medium confidence*). Because the ocean has a higher heat capacity than the atmosphere (heat  
31 capacity of the upper 100 m of the ocean is about 30 times larger than the heat capacity of the atmosphere),  
32 the partitioning of heat between the atmosphere and the ocean is primarily influenced by the temperature  
33 differences between air and seawater. Similarly, the unique seawater carbonate buffering capacity, enables  
34 CO<sub>2</sub> to be stored in the ocean as dissolved salts, rather than just as dissolved gas, which increases the  
35 capacity of seawater to store CO<sub>2</sub> by two orders of magnitude beyond the solubility of CO<sub>2</sub> gas and  
36 approximate the partitioning ratio of heat between the atmosphere and the ocean (Zeebe and Wolf-Gladrow,  
37 2009; Bronselaer and Zanna, 2020; Section 9.2.2.1). The role of the biological carbon pump in influencing  
38 the ocean sink of anthropogenic carbon into the ocean interior is assessed to be minimal during the historical  
39 period but this may change, particularly in regional contexts, by 2100 (*medium confidence*) (Laufkötter et al.,  
40 2015; Kwiatkowski et al., 2020). Its role is important in the natural or pre-industrial carbon cycle (*medium*  
41 *confidence*) (Henson et al., 2016).

42  
43 Under climate change, the buffering capacity of the ocean decreases (increasing Revelle Factor), which  
44 reflects a decreasing capacity for the ocean to take up additional anthropogenic CO<sub>2</sub> and store it in the DIC  
45 reservoir (Egleston et al., 2010). In contrast to CO<sub>2</sub>, there is no physical limitation that would reduce the  
46 ability of surface ocean temperature to equilibrate with the atmospheric temperature. However, both carbon  
47 and heat fluxes depend on air-sea heat fluxes that depend on gradients of characteristics at the air-sea  
48 interface. These gradients at the air-sea interface respond to ocean dynamics, such as the volume of the  
49 surface mixed-layer that equilibrates with the atmosphere, and ocean circulation that can flush the surface  
50 layer with water-masses that have not equilibrated with the atmosphere for a long time. Limited recent  
51 evidence has suggested that the effect of small-scale dynamics absent in climate and earth system models  
52 might be locally important in this regard (Bachman and Klocker, 2020). In summary, changes in heat and  
53 carbon uptake by the ocean rely on a combination of unique chemical and shared physical processes any of  
54 which have the potential to disrupt the coherence of heat and CO<sub>2</sub> change in the ocean.

### *Spatial pattern of air-sea fluxes and storage*

Large scale regional and global ocean circulation shape the spatial pattern of the uptake and storage of both CO<sub>2</sub> and heat (see Figure 5.8 for carbon; Figure 9.6 for heat observations; Section 9.2; (Frölicher et al., 2015; Bronselaer and Zanna, 2020). This coherence of spatial patterns driven by the large-scale ocean circulation has three aspects: firstly, notwithstanding interannual - decadal variability in heat and CO<sub>2</sub> uptake, there is a spatial coherence of the temporally integrated uptake at the air-sea boundary, particularly in the Southern Ocean (Cross-Chapter Box 5.3, Figure 1; Talley et al., 2016; Keppler and Landschützer, 2019; Auger et al., 2021). Secondly, the importance of the meridional overturning circulation in the subsequent storage of both heat and CO<sub>2</sub> in mode, intermediate and deep waters of the ocean interior (Section 9.2). Of particular note are the roles of the North Atlantic Ocean (Section 9.2.3.1) and the Southern Ocean (Section 9.2.3.2) in linking the spatial pattern of air-sea fluxes, the storage of heat and carbon, and ultimately in understanding and predicting the sensitivity of the carbon – heat nexus to climate change (Frölicher et al., 2015; Thomas et al., 2018; Wu et al., 2019).

**[START CROSS-CHAPTER BOX 5.3, FIGURE 1 HERE]**

**Cross-Chapter Box 5.3, Figure 1: CMIP6 multi-model mean of changes in zonally integrated (a) carbon and (b) heat storage in ocean between the pre-industrial and the modern period.**

Carbon corresponds to dissolved inorganic carbon. Data are shown for the upper 2000m. The modern period is 1995–2014. Adapted from (Frölicher et al., 2015))

**[END CROSS-CHAPTER BOX 5.3, FIGURE 1 HERE]**

The role of the large-scale circulation in shaping these fluxes is to (i) flush the ocean surface layer with deep waters that are relatively cold and with weak or no anthropogenic CO<sub>2</sub> and heat content because they have been isolated from the atmosphere for centuries, and (ii) transport the anthropogenic CO<sub>2</sub> and heat at depth, away from the atmosphere (Frölicher et al., 2015; Marshall et al., 2015; Armour et al., 2016). For instance, in the Southern Ocean upwelled water-masses uptake large amount of anthropogenic CO<sub>2</sub> and heat (Cross-Chapter Box 5.3, Figure 1), which are then exported northward by the circulation to be stored at depth in the Southern Hemisphere subtropical gyres (Cross-Chapter Box 5.3, Figure 1; Figure 9.7). In the North Atlantic, the signature of the Atlantic meridional overturning circulation (AMOC) is also clearly visible, with large amounts of heat and carbon being stored beneath the North Atlantic subtropical gyre at 1 km depth (Cross-Chapter Box 5.3, Figure 1). In summary, the net air-sea fluxes of anthropogenic CO<sub>2</sub> and heat depend on large-scale circulation, which is associated with upper ocean stratification, mixed-layer depth, and water-mass formation, transport and mixing (Sections 9.1; 9.2; 9.3).

### *Changes in ocean processes and impact on the ocean carbon-heat nexus*

Future projections of the ocean carbon-heat nexus in the second half of the 21st century, particularly those under weak or no mitigation scenarios, are characterized by the strengthening of the two largest positive feedbacks: weakening surface ocean CO<sub>2</sub> buffering capacity (increasing Revelle Factor) and warming that further reduces CO<sub>2</sub> solubility and strengthens ocean stratification, which reduces exchange between the ocean surface and interior (Jiang et al., 2019; Bronselaer and Zanna, 2020). These are offset by a growing but scenario-dependent negative feedback from increasing carbon and heat air sea fluxes towards the ocean, due to increased atmospheric temperature and CO<sub>2</sub> concentrations (Talley et al., 2016; Jiang et al., 2019; McKinley et al., 2020). The Southern Ocean (SO) in particular is one of the regions where the projected feedback can be largest and where inter-model differences are strongest (Roy et al., 2011; Frölicher et al., 2015; Hewitt et al., 2016; Mongwe et al., 2018). These projected trends in ocean carbonate chemistry (5.4.2), together with surface ocean warming (Section 9.2.1.1), explain the slow down and long-term reduction of the ocean sink for anthropogenic CO<sub>2</sub> even as emissions continue to rise beyond 2050 under weak to no mitigation scenarios (Section 5.4: Figure 5.25; Technical Summary TS Box 7 and Figure 2.7.1). Projected change in the North Atlantic and Southern Ocean overturning circulation also impact air-sea fluxes of heat

1 and carbon. The *very likely* decline in AMOC in the 21st century for all shared socioeconomic pathways  
2 (SSP) scenarios (Section 9.2.3.1) tends to reduce heat and carbon uptake translating in a positive feedback.  
3 In contrast, in the Southern Ocean, the future 21st century projected increase in upper ocean overturning  
4 circulation (*low confidence*) due to increasing wind forcing projected for all scenarios, except those with  
5 large mitigation (SSP1–2.6), translate in a negative feedback with increasing heat and carbon uptake and  
6 storage despite the increasing stratification and outgassing of natural CO<sub>2</sub> in the upwelling zone (Sections  
7 9.2.3.2; 5.2.1.3).

8  
9 In summary, a combination of unique chemical properties of seawater carbonate combined with shared  
10 physical ocean processes explain the coherence and scaling in the uptake and storage of both CO<sub>2</sub> and heat in  
11 the ocean, which is the basis for the carbon-heat nexus (*high confidence*). In this way the processes of the  
12 ocean carbon – heat nexus help understand the quasi-linear and path independence of properties of TCRE,  
13 which forms the basis for the ZEC (Section 5.5) (*medium confidence*). Future projections under low or no  
14 mitigation indicate with *high confidence* that carbon chemistry and warming will strengthen the positive  
15 feedback to climate change by reducing ocean carbon uptake, and *medium confidence* that ocean circulation  
16 may partially compensate that positive feedback by slightly increasing anthropogenic carbon storage.  
17 Increasing ocean warming and stratification may decrease exchanges between the surface and subsurface  
18 ocean, which could reduce the path independence of TCRE, though this effect can be partially  
19 counterbalanced regionally by increasing circulation associated with increasing winds (*low confidence*).

20  
21 **[END CROSS-CHAPTER BOX 5.3]**

#### 22 23 24 5.5.1.2 *Assessment of Limits of the TCRE Concept*

##### 25 26 5.5.1.2.1 *Sensitivity to amount of cumulative CO<sub>2</sub> emissions*

27 AR5 indicated that the concept of a constant ratio of cumulative emissions of CO<sub>2</sub> to temperature was  
28 applicable to scenarios with increasing cumulative CO<sub>2</sub> emissions up to 2000 PgC (Collins et al., 2013a).  
29 Recent analyses have added confidence to this insight (Herrington and Zickfeld, 2014; Steinacher and Joos,  
30 2016) and have also shown some evidence of a potentially larger window of constant TCRE (Leduc et al.,  
31 2015; Tokarska et al., 2016). Using an analytical approach, MacDougall and Friedlingstein (2015) quantified  
32 a window of constant TCRE – defined as the range in cumulative emissions over which the TCRE remains  
33 within 95% of its maximum value – as between 360 to 1560 PgC. However, models with a more  
34 sophisticated ocean representation suggest that TCRE could also remain constant for considerably larger  
35 quantities of cumulative emissions, up to at least 3000 PgC (Leduc et al., 2015; Tokarska et al., 2016).  
36 Beyond this upper limit, studies are inconclusive with some suggesting that TCRE will decrease (Leduc et  
37 al., 2015) and others indicating that the linearity would hold up to as much as 5000 PgC (Tokarska et al.,  
38 2016).

39  
40 As cumulative emissions increase, weakened land and ocean carbon sinks increase the airborne fraction of  
41 CO<sub>2</sub> emissions (see Section 5.4.5, Figure 5.25), but each unit increase in atmospheric CO<sub>2</sub> has a smaller  
42 effect on global temperature owing to the logarithmic relationship between CO<sub>2</sub> and its radiative forcing  
43 (Matthews et al., 2009; Etminan et al., 2016). At high values of cumulative emissions, some models simulate  
44 less warming per unit CO<sub>2</sub> emitted, suggesting that the saturation of CO<sub>2</sub> radiative forcing becomes more  
45 important than the effect of weakened carbon sinks (Herrington and Zickfeld, 2014; Leduc et al., 2015). The  
46 behaviour of carbon sinks at high emissions levels remains uncertain, as models used to assess the limits of  
47 the TCRE show a large spread in net land carbon balance (Section 5.4.5) and most estimates did not include  
48 the effect of permafrost carbon feedbacks (Sections 5.5.1.2.3, 5.4). The latter would tend to further increase  
49 the airborne fraction at high cumulative emissions levels, and could therefore extend the window of linearity  
50 to higher total amounts of emissions (MacDougall et al., 2015). Leduc et al. (2016) suggested further that a  
51 declining strength of snow and sea-ice feedbacks in a warmer world would also contribute to a smaller  
52 TCRE at high amounts of cumulative emissions. On the other hand, Tokarska et al. (2016) suggested that a  
53 large decrease in TCRE for high cumulative emissions is only associated with some EMICs; in the four  
54 ESMs analysed in their study, the TCRE remained approximately constant up to 5000 PgC, owing to  
55 stronger declines in the efficiency of ocean heat uptake in ESMs compared to EMICs.

1  
2 Overall, there is *high agreement* between multiple lines of evidence (*robust evidence*) resulting in *high*  
3 *confidence* that TCRE remains constant for the domain of increasing cumulative CO<sub>2</sub> emissions until at least  
4 1500 PgC, with *medium confidence* of it remaining constant up to 3000 PgC because of less agreement  
5 across available lines of evidence.  
6  
7

#### 8 5.5.1.2.2 *Sensitivity to the Rate of CO<sub>2</sub> Emissions*

9 Global average temperature increase responds over a timescale of about 10 years following the emission of a  
10 100 PgC pulse of CO<sub>2</sub> (Joos et al., 2013; Ricke and Caldeira, 2014), with larger emission pulses associated  
11 with longer timescales and smaller pulses with shorter ones (Joos et al., 2013; Matthews and Solomon, 2013;  
12 Zickfeld and Herrington, 2015). This behaviour is confirmed in other studies, including those that calculate  
13 the temperature response to an instantaneous doubling or quadrupling of atmospheric CO<sub>2</sub> (Matthews et al.,  
14 2009; Gillett et al., 2013; Herrington and Zickfeld, 2014; Leduc et al., 2015; Hajima et al., 2020b). These  
15 findings suggest that the TCRE is sensitive to the rate of emissions, but studies assessing this sensitivity have  
16 found diverging results. For example, an increase in TCRE and its surrounding uncertainty was reported for  
17 experiments that imply a gradual decline in annual CO<sub>2</sub> emissions (Tachiiri et al., 2019). These studies  
18 suggest that in most cases TCRE would be expected to increase in scenarios with decreasing annual  
19 emissions rates. This increase in TCRE for annual CO<sub>2</sub> emissions declining towards zero can be the result of  
20 the zero emissions commitment (ZEC) which is the amount of warming projected to occur following a  
21 complete cessation of emissions (see Chapter 4, Section 4.7.2 for its assessment), as well as Earth system  
22 processes that are unrepresented in current TCRE estimates (Section 5.5.2.2.4) and other factors. When using  
23 TCRE to estimate CO<sub>2</sub> emissions consistent with a specific maximum warming level these factors have to be  
24 taken into account (see Figure 5.31). Combined with recent literature on the ZEC (MacDougall et al., 2020)  
25 and emissions pathways (Huppmann et al., 2018) and noting the lack of literature that disentangles these  
26 various contributions, there is *medium evidence* and *high agreement* resulting in *medium confidence* that  
27 TCRE remains a good predictor of CO<sub>2</sub>-induced warming when applied in the context of emission reduction  
28 pathways, provided that ZEC and long-term Earth system feedbacks are adequately accounted for when  
29 emissions decline towards zero (see also next section).  
30  
31

#### 32 5.5.1.2.3 *Reversibility and Earth System Feedbacks*

33 There are relatively few studies that have assessed how the TCRE is expected to change in scenarios of  
34 declining emissions followed by net negative annual CO<sub>2</sub> emissions. Conceptually, the literature suggests  
35 that the small lag of about a decade between CO<sub>2</sub> emissions and temperature change (Ricke and Caldeira,  
36 2014; Zickfeld and Herrington, 2015) would result in more warming at a given amount of cumulative  
37 emissions in a scenario where that emission level is first exceeded and then returned to by deploying  
38 negative emissions (referred to as an “overshoot”, as is often the case in scenarios that aim to limit radiative  
39 forcing in 2100 to 2.6 or 1.9 W/m<sup>2</sup> (Riahi et al., 2017; Rogelj et al., 2018a). Zickfeld et al. (2016) showed  
40 this to hold across a range of scenarios with positive emissions followed by negative emissions, whereby the  
41 TCRE increased by about 10% across the transition from positive to negative emissions as a result of the  
42 thermal and carbon inertia of the deep ocean. However, CMIP6 results for the SSP5–3.4-overshoot scenario  
43 show diverging trends across various ESMs (see Section 5.4.10, Figure 5.30). In an idealised CO<sub>2</sub>-  
44 concentration-driven setting Tachiiri et al. (2019) also reported an increase in TCRE. Exploring pathways  
45 with emissions rates and overshoots closer to mitigation pathways considered over the 21st century (in this  
46 case up to about 300 PgC), a recent emission-driven EMIC experiment showed pathway independence of  
47 TCRE (Tokarska et al., 2019b). Furthermore, also in absence of net negative emissions, warming would not  
48 necessarily remain perfectly constant on timescales of centuries and millennia, but could both decrease or  
49 increase (Frölicher and Paynter, 2015; Williams et al., 2017b; Hajima et al., 2020b). These additional  
50 changes in global mean temperature increase at various timescales are known as the ZEC (Jones et al., 2019;  
51 MacDougall et al., 2020), assessed in Section 4.7.2, and have to be integrated when using TCRE to estimate  
52 warming or remaining carbon budgets in overshoot scenarios.  
53

54 The AR5-assessed (Collins et al., 2013b) TCRE range was based in part on the ESMs available at the time,  
55 which did not include some potentially important Earth system feedbacks. Since then, a number of studies

1 have assessed the importance of permafrost carbon feedbacks in particular on remaining carbon budgets  
2 (MacDougall et al., 2015; MacDougall and Friedlingstein, 2015; Burke et al., 2017; Gasser et al., 2018;  
3 Lowe and Bernie, 2018), a development highlighted and assessed in the IPCC Special Report on Global  
4 Warming of 1.5°C (Rogelj et al., 2018b). MacDougall and Friedlingstein (2015) reported a TCRE increase  
5 of about 15% when including permafrost carbon feedbacks. The overall linearity of the TCRE during the  
6 21st century was not affected, but they also found that permafrost carbon feedbacks caused an increase in  
7 TCRE on multi-century timescales under declining CO<sub>2</sub> emission rates. In addition, other processes that are  
8 not or only partially considered in individual or all ESMs could cause a further increase or decrease of TCRE  
9 (Matthews et al., 2020). These are discussed in detail in Section 5.4, but their quantitative effects on TCRE  
10 have not yet been explored by the literature.

11  
12 Whether TCRE remains an accurate predictor of CO<sub>2</sub>-induced warming when annual CO<sub>2</sub> emissions reach  
13 zero and are followed by net carbon-dioxide removal (also referred to as TCRE reversibility) therefore  
14 hinges on contributions of slow components of the climate system that cause unrealised warming from past  
15 CO<sub>2</sub> emissions. Such slow components can arise from either physical climate (i.e., ocean heat uptake) or  
16 carbon cycle (i.e., ocean carbon uptake and permafrost carbon release) processes. The combined effect of  
17 these processes determines the magnitude and sign of the ZEC (MacDougall et al., 2020), which in turn  
18 impacts TCRE reversibility. As discussed in Section 4.7.2, recent model estimates of the ZEC suggest a  
19 range of  $\pm 0.19^{\circ}\text{C}$  centred on zero (MacDougall et al., 2020). This suggests *limited agreement* among models  
20 as to the reversibility of the TCRE in response to net-negative CO<sub>2</sub> emissions. Furthermore, most models  
21 used for ZEC assessments to date do not represent permafrost carbon processes, although understanding their  
22 contribution is essential to isolate the TCRE contribution. There is therefore *limited evidence* that quantifies  
23 the impact of permafrost carbon feedbacks on the reversibility of TCRE, leading to *low confidence* that the  
24 TCRE remains an accurate predictor of temperature changes in scenarios of net-negative CO<sub>2</sub> emissions on  
25 timescales of more than a half a century.

### 26 27 28 5.5.1.3 Estimates of TCRE

29  
30 IPCC AR5 (Collins et al., 2013a) assessed TCRE *likely* to fall in the range of 0.8–2.5°C per 1000 PgC (or  
31 per exagrams of carbon, EgC<sup>-1</sup>) for cumulative emissions up to 2000 PgC, based on multiple lines of  
32 evidence. These include estimates based on Earth system models of varying complexity (Matthews et al.,  
33 2009; Gillett et al., 2013; Zickfeld et al., 2013), simple climate modelling approaches (Allen et al., 2009;  
34 Rogelj et al., 2012) or observational constraints and attributable warming (Gillett et al., 2013).

35  
36 Since IPCC AR5, new studies have further expanded the evidence base for estimating the value of TCRE.  
37 These studies rely on ESMs or EMICs, observational constraints and concepts of attributable warming, or  
38 theoretically derived equations (see Table 5.7 for an overview). Several studies have endeavoured to  
39 partition the uncertainty in the value of TCRE into constituent sources. For example, TCRE can be  
40 decomposed into terms of TCR and the airborne fraction of anthropogenic CO<sub>2</sub> emissions over time (Allen et  
41 al., 2009; Matthews et al., 2009). These two terms are assessed individually (see Section 5.4 and Chapter 7,  
42 respectively) and allow the integration of evidence assessed elsewhere in the report into the assessment of  
43 TCRE (Section 5.5.1.4). Further studies use a variety of methods including analysing CMIP5 (Williams et  
44 al., 2017c) or CMIP6 (Arora et al., 2020; Jones and Friedlingstein, 2020) output, conducting perturbed  
45 parameter experiments with a single model (MacDougall et al., 2017), Monte-Carlo methods applied to a  
46 simple climate model (Spafford and Macdougall, 2020), or observations and estimates of the contribution of  
47 CO<sub>2</sub> and non-CO<sub>2</sub> forcings (Matthews et al., 2021). All of the studies agree that uncertainty in climate  
48 sensitivity (either Equilibrium Climate Sensitivity (ECS) or Transient Climate Response (TCR)) is amongst  
49 the most important contribution to uncertainty in TCRE, with uncertainty in the strength of the land carbon  
50 feedback and ocean heat uptake or ventilation having also been identified as crucial to uncertainty in TCRE  
51 (Matthews et al., 2009; Gillett et al., 2013; Ehlert et al., 2017; MacDougall et al., 2017; Williams et al.,  
52 2017b, 2020; Katavouta et al., 2019; Arora et al., 2020; Jones and Friedlingstein, 2020; Spafford and  
53 Macdougall, 2020). Finally, internal variability has been shown to affect the maximum accuracy of TCRE  
54 estimates by  $\pm 0.1^{\circ}\text{C}$  per 1000 PgC (5–95% range) (Tokarska et al., 2020).



## [START TABLE 5.7 HERE]

**Table 5.7: Overview of estimates of studies estimating the transient response to cumulative emissions (TCRE) of CO<sub>2</sub>.** GSAT = Global mean surface air temperature increase, SAT = surface air temperature (e.g. over land only), SST = sea surface temperature, ECS = equilibrium climate sensitivity. Studies that do not isolate the CO<sub>2</sub>-induced warming contribution in their TCRE estimates are not included.

Study	TCRE Range (°C per 1000 PgC)	Notes
<b>Studies available at the time of IPCC AR5</b>		
(Matthews et al., 2009)	1–2.1	5 to 95% range; GSAT; C <sup>4</sup> MIP model range
(Allen et al., 2009)	1.4–2.5	5 to 95% range; blended global mean SAT and SSTs (no infilling of coverage gaps); simple model
(Zickfeld et al., 2009)	1.5	Best estimate; GSAT, EMIC
(Williams et al., 2012)	0.8–1.9	Range consistent with 2 to 4.5 °C ECS; GSAT
(Rogelj et al., 2012)	About 1–2	5 to 95% range; historical constraint on GMST increase, but other constraints on GSAT increase MAGICC model calibrated to C <sup>4</sup> MIP model range and 2–4.5°C <i>likely</i> ECS
(Zickfeld et al., 2013)	1.4–2.5; mean: 1.9	Model range; GSAT, EMICs
(Eby et al., 2013)	1.1–2.1; mean: 1.6	Model range; GSAT, EMICs
(Gillett et al., 2013)	0.8–2.4	Model range; GSAT, CMIP5 ESMs
(Gillett et al., 2013)	0.7–2.0	5 to 95% range; blended global mean SAT and SSTs; observationally constrained estimates of historical warming and emissions
IPCC AR5 (Collins et al., 2013a)	0.8–2.5	Assessed <i>likely</i> range; multiple lines of evidence; mixed definition of global average temperature increase
<b>Studies published since IPCC AR5</b>		
(Tachiiri et al., 2015)	0.3–2.4	5 to 95% range; blended global mean SAT and SSTs; JUMP-LCM model perturbed physics ensemble (EMIC)
(Tachiiri et al., 2015)	1.1–1.7	5 to 95% range; blended global mean SAT and SSTs; observationally constrained JUMP-LCM perturbed physics ensemble
(Goodwin et al., 2015)	1.1 ± 0.5	5 to 95% range; theoretically derived TCRE equation constrained by surface warming, radiative forcing, and historic ocean and land carbon uptake from IPCC AR5
(Millar et al., 2017b)	1.0–2.5	5 to 95% range; blended global mean SAT and SSTs (HadCRUT4); observationally constrained probabilistic setup of simple climate model
(Steinacher and Joos, 2016)	1.0–2.7; median: 1.7	5 to 95% range; GSAT, observationally constrained BERN3D-LPJ EMIC
(MacDougall et al., 2017)	0.9–2.5; mean: 1.7	5 to 95% range; GSAT, emulation of 23 CMIP5 ESMs
(Ehlert et al., 2017)	1.2–2.1	Model range; GSAT, UVIC EMIC with varying ocean mixing parameters
(Williams et al., 2017c)	1.4–2.1; mean: 1.8	1-sigma range; GSAT, diagnosed from 10 CMIP5 ESMs
(Millar and Friedlingstein, 2018)	0.9–2.6; best estimate: 1.3	5 to 95% range; blended global mean SAT and SSTs (Cowtan and Way, 2014); detection attribution with observational constraints
(Millar and Friedlingstein, 2018)	best estimate: 1.5	Blended global mean SAT and SSTs (Berkeley Earth); detection attribution with observational constraints
(Millar and Friedlingstein, 2018)	best estimate: 1.2	Blended global mean SAT and SSTs (Cowtan and Way, 2014); detection attribution with observational constraints, with updated historical CO <sub>2</sub> emissions (Le Quéré et al., 2018b)
(Smith et al., 2018a)	1.0–2.2	5 to 95% range; blended global mean SAT and SSTs (Cowtan and Way, 2014); observationally constrained probabilistic setup of simple climate model
(Matthews et al., 2021)	1.0–2.2; median: 1.5	5 to 95% range; blended global mean SAT and SSTs; human-induced warming (Haustein et al., 2017) based on an average of three full coverage datasets; observationally constrained estimate using the current non-CO <sub>2</sub> fraction of total anthropogenic forcing
(Arora et al., 2020)	1.3–2.4; mean: 1.8; median: 1.65	Model range; GSAT, diagnosed CO <sub>2</sub> emissions in CMIP6 ESMs
(Williams et al., 2020)	1.2–2.1; mean: 1.6	1-sigma range; GSAT, diagnosed CO <sub>2</sub> emissions in 9 CMIP6 ESMs
(Jones and Friedlingstein, 2020)	1.2–2.7; median: 1.8	5 to 95% range; GSAT; Estimate based on decomposition presented in (Jones and Friedlingstein, 2020) with ranges of carbon cycle feedback parameters from CMIP6 (Arora et al., 2020), see Section 5.4.
(Spafford and Macdougall, 2020)	1.1–2.9; mean: 1.9; median: 1.8	5 to 95% range; ratio of land SAT and SST; probabilistic assessment of with a zero-dimensional ocean diffusive model
<b>Cross-AR6 lines of evidence</b>		
Transient Climate Response (TCR) and Airborne	1.0–2.3; median: 1.6	5 to 95% range; GSAT; TCR–AF decomposition-based estimate using the assessed range of TCR (Section 7.5, 1.8°C median with 0.4°C 1-sigma

Fraction (AF)		range) and an airborne fraction of $53 \pm 6\%$ (1-sigma range)
<b>Overall assessment</b>		
IPCC AR6	1.0–2.3; best estimate: 1.65	<i>Likely</i> range; GSAT; Based on combination of cross-AR6 lines of evidence (Section 5.5.1.4); normally distributed

[END TABLE 5.7 HERE]

#### 5.5.1.4 Combined assessment of TCRE

Studies differ in how they define TCRE, in the methods they use, and their assumptions, such as the assumed climate sensitivity distribution or the choice of metrics of global temperature change (e.g. GMST or GSAT, see Table 5.7). This makes TCRE estimates from individual studies difficult to compare. The combined assessment of TCRE therefore takes advantage of the well-established decomposition of TCRE in two factors: the transient climate response (TCR) and the airborne fraction (Section 5.5.1.3). This provides a TCRE assessment range for CO<sub>2</sub>-induced warming at the time of doubling CO<sub>2</sub> concentrations that builds on the broader Working Group 1 assessment. Expert judgment based on the airborne fraction range found in CMIP6 models (Arora et al., 2020; Jones and Friedlingstein, 2020) suggest a value of 53% with a 1-sigma range of  $\pm 6\%$ , which is double the sigma range based on the spread of CMIP6 models only. Combining this range with the AR6 TCR assessment (Section 7.5, best estimate 1.8°C, 1.4–2.2°C *likely* and 1.2–2.4°C *very likely* range) results in a 5–95% range of 1.0–2.3°C per 1000 PgC. Based on expert judgment that accounts for the incomplete coverage of all Earth system components, this result in a consolidated assessment that TCRE would fall *likely* in the range of 1.0–2.3°C per 1000 PgC, with a best estimate of 1.65°C per 1000 PgC. Warming here reflects the human-induced GSAT increase and assumes a normal distribution. Some studies using observational constraints support a lognormal shape for the TCRE distribution (Spafford and Macdougall, 2020), but such a distribution is currently not supported by the combined assessment of TCR and airborne fraction. Finally, this assessed TCRE range needs to be considered in combination with the ZEC (Section 4.7.2) when estimating the CO<sub>2</sub>-induced warming of low-emissions scenarios.

#### 5.5.2 Remaining Carbon Budget Assessment

Estimates of remaining carbon budgets consistent with holding global warming below a specific temperature threshold depend on a range of factors which are increasingly being studied and quantified. These factors include (i) well-understood methodological and definitional choices (Friedlingstein et al., 2014a; Rogelj et al., 2016, 2018b) (see Sections 5.5.2.1 and Section 5.5.2.2), and (ii) a set of contributing factors such as historical warming, the TCRE and its limitations, the ZEC (the amount of warming projected to occur following a complete cessation of emissions, see Section 4.7.2), as well as contributions of non-CO<sub>2</sub> climate forcings (Section 5.5.2.2) (Rogelj et al., 2015a, 2015b; MacDougall and Friedlingstein, 2015; Simmons and Matthews, 2016; MacDougall, 2016; Ehlert et al., 2017; Matthews et al., 2017, 2021; Millar et al., 2017a; Tokarska et al., 2018; Goodwin et al., 2018; Mengis et al., 2018; Pfliegerer et al., 2018; Cain et al., 2019). These contributing factors are integrated in an overarching assessment of remaining carbon budgets for limiting global average warming to levels ranging from 1.5°C to 2.5°C relative to pre-industrial levels provided in Section 5.5.2.3. Box 5.2 provides an overview of the methodological advances since AR5 (Collins et al., 2013b).

##### 5.5.2.1 Framework and Earlier Approaches

The IPCC AR6 Glossary (Annex VII) defines remaining carbon budgets as the maximum amount of cumulative net global anthropogenic CO<sub>2</sub> emissions expressed from a recent specified date that would result in limiting global warming to a given level with a given probability, taking into account the effect of other anthropogenic climate forcings, consistent with their assessment in the IPCC SR1.5 (Rogelj et al., 2018b). Studies, however, apply a variety of definitions that result in published remaining carbon budget estimates informing to cumulative emissions at the time when global-mean temperature increase would reach, exceed,

1 avoid, or peak at a given warming level with a given probability, for example (Collins et al., 2013a; Stocker  
2 et al., 2013c; Clarke et al., 2014; Friedlingstein et al., 2014a; IPCC, 2014; Rogelj et al., 2016; Millar et al.,  
3 2017a). This section provides an assessment of remaining carbon budgets consistent with the AR6 Glossary  
4 definition (Annex VII). Given that some feedbacks are time dependent, the values in this section apply to  
5 limiting warming over the 21st century, consistent with recent studies highlighting the usefulness of time-  
6 limited carbon budgets (Sanderson, 2020). Irrespective of the exact definition of the remaining carbon  
7 budget, the finding that higher cumulative CO<sub>2</sub> emissions lead to higher temperatures implies that annual net  
8 CO<sub>2</sub> emissions have to decline to close to zero in order to halt global warming, whether at 1.5°C, 2°C or  
9 another level (Allen et al., 2018).

10  
11 Two approaches were used in AR5 to determine carbon budgets (Collins et al., 2013; Stocker et al., 2013;  
12 Clarke et al., 2014; IPCC, 2014; Rogelj et al., 2016). Working group I (WGI) reported threshold exceedance  
13 budgets (TEB) that correspond to the amount of cumulative CO<sub>2</sub> emissions at the time a specific temperature  
14 threshold is exceeded with a given probability in a particular greenhouse-gas and aerosol (pre-cursor)  
15 emission scenario (Collins et al., 2013a; IPCC, 2013b; Stocker et al., 2013c). WGI also reported TEBs for  
16 the hypothetical case that only CO<sub>2</sub> would be emitted by human activities (Collins et al., 2013a; IPCC,  
17 2013b; Stocker et al., 2013c). AR5 Working group III used threshold avoidance budgets (TAB) that  
18 correspond to the cumulative CO<sub>2</sub> emissions over a given time period of a subset of greenhouse-gas and  
19 aerosol (precursor) emission scenarios in which global-mean temperature increase stays below a specific  
20 temperature threshold with at least a given probability (Clarke et al., 2014). The AR5 synthesis report used  
21 TABs defined until the time of peak warming over the 21st century (IPCC, 2014). Drawbacks have been  
22 identified for both TEBs and TABs (Rogelj et al., 2016). TABs provide an estimate of the cumulative CO<sub>2</sub>  
23 emissions under pathways that have as a common characteristic that they do not exceed a specific global  
24 warming threshold. The actual level of maximum warming can however vary between pathways, leading to  
25 an unnecessary and poorly constrained spread in TAB estimates (Rogelj et al., 2016). The TAB approach  
26 does therefore typically not result in accurate projections of the remaining carbon budget. On the other hand,  
27 a drawback of TEBs is that they provide an estimate of the cumulative CO<sub>2</sub> emissions at the time global  
28 warming crosses a given threshold of interest in a specific emissions scenario, for example, most of the  
29 standard scenarios used in climate change research such as the RCPs or SSP-based scenarios exceed global  
30 warming of 1.5°C or 2°C (see Cross-Chapter Box 1.5) (Collins et al., 2013a; Stocker et al., 2013c;  
31 Friedlingstein et al., 2014a; Millar et al., 2017a). Because of potential variations in non-CO<sub>2</sub> warming at that  
32 point in time or potential lags of about a decade in CO<sub>2</sub> warming (Joos et al., 2013; Ricke and Caldeira,  
33 2014; Zickfeld and Herrington, 2015; Rogelj et al., 2015a, 2016, 2018) TEBs also do not provide a precise  
34 estimate of the remaining carbon budget for limiting warming to a specific level.

35  
36 Since the publication of AR5 (Collins et al., 2013b), several new approaches have been proposed that  
37 provide a solution to the identified limitations of TABs and TEBs. Most of these approaches indirectly rely  
38 on the concept of TCRE (Section 5.5.1), for example, because they estimate modelled cumulative CO<sub>2</sub>  
39 emissions until a temperature threshold is crossed and use this budget to infer insights for pathways which  
40 attempt to limit warming to below this threshold and thus need to follow a different path (Friedlingstein et  
41 al., 2014; Matthews et al., 2017; Millar et al., 2017; Goodwin et al., 2018; Tokarska and Gillett, 2018). In  
42 this report, the assessment framework of the IPCC SR1.5 for remaining carbon budgets is applied (Rogelj et  
43 al., 2018b, 2019). This framework allows to integrate multiple lines of evidence to assess the contributions of  
44 five components that together result in a consolidated assessment of the remaining carbon budget (historical  
45 warming, TCRE, non-CO<sub>2</sub> warming, the ZEC, and adjustments due to additional Earth system feedbacks, see  
46 Section 5.5.2.2). It builds on the advances in estimating remaining carbon budgets or related quantities that  
47 have been published since AR5 (Rogelj et al., 2015a; Haustein et al., 2017; Matthews et al., 2017, 2021;  
48 Millar et al., 2017a; Gasser et al., 2018; Lowe and Bernie, 2018; Tokarska et al., 2018; Nicholls et al., 2020).

49  
50 Recent studies suggest further changes to this framework by including non-linear adjustments to the TCRE  
51 contribution (Nicholls et al., 2020), or including non-CO<sub>2</sub> forcers in different ways by accounting for their  
52 different forcing effects (Matthews et al., 2021). Figure 5.31 provides a conceptual schematic of how the  
53 various individually assessed contributions are combined into a consolidated assessment of the remaining  
54 carbon budget. Together with estimates of historical CO<sub>2</sub> emissions to date (Section 5.2.1), these remaining  
55 carbon budgets provide the overall amount of cumulative CO<sub>2</sub> emissions consistent with limiting global

warming to specific levels. A comparison with the approach applied in AR5 (Collins et al., 2013; Clarke et al., 2014) is available in the IPCC SR1.5 Section 2.2.2 (Rogelj et al., 2018b) as well as Box 5.2.

[START FIGURE 5.31 HERE]

**Figure 5.31: Illustration of relationship between cumulative emissions of carbon dioxide (CO<sub>2</sub>) and global mean surface air temperature increase (left) and conceptual schematic of the assessment of the remaining carbon budget from its constituting components (right).** Carbon budgets consistent with various levels of additional warming are provided in Table 5.8 and should not be read from the illustrations in either panel. Left-hand panel: Historical data (thin black line data) shows historical CO<sub>2</sub> emissions as reported in (Friedlingstein et al., 2020) together with the assessed global mean surface air temperature increase from 1850–1900 as assessed in Chapter 2 (Box 2.3, GSAT). The orange-brown range with its central line shows the estimated human-induced share of historical warming (Haustein et al., 2017). The vertical orange-brown line shows the assessed range of historical human-induced warming for the 2010–2019 period relative to 1850–1900 (Chapter 3). The grey cone shows the assessed range for the transient climate response to cumulative emissions of carbon dioxide (TCRE) assessed to fall *likely* in the 1.0–2.3 °C per 1000 PgC range (Section 5.5.1.4), starting from 2015. Thin coloured lines show CMIP6 simulations for the five scenarios of the AR6 core set (SSP1–1.9, green; SSP1–2.6, blue; SSP2–4.5, yellow; SSP3–7.0, red; SSP5–8.5, maroon), starting from 2015. Diagnosed carbon emissions (Arora et al., 2020) are complemented with estimated land-use change emissions for each respective scenario (Gidden et al., 2018). Coloured areas show the Chapter 4 assessed *very likely* range of GSAT projections and thick coloured central lines the median estimate, for each respective scenario, relative to the original scenario emissions (Riahi et al., 2017; Gidden et al., 2018; Rogelj et al., 2018a). Right-hand panel: schematic illustration of assessment of remaining carbon budget based on multiple lines of evidence. The remaining allowable warming is estimated by combining the global warming limit of interest with the assessed historical human induced warming (Section 5.5.2.2.2), the assessed future potential non-CO<sub>2</sub> warming contribution (Section 5.5.2.2.3) and the ZEC (Section 5.5.2.2.4). Note that contributions in the right-hand panel are illustrative and contributions are not to scale. For example, the central ZEC estimate was assessed to be zero. The remaining allowable warming (vertical blue bar) is subsequently combined with the assessed TCRE (Sections 5.5.1.4 and 5.5.2.2.1) and contribution of unrepresented Earth system feedbacks in models used to estimate ZEC and TCRE (Section 5.5.2.2.5) to provide an assessed estimate of the remaining carbon budget (horizontal blue bar, Table 5.8). Further details on data sources and processing are available in the chapter data table (Table 5.SM.6).

[END FIGURE 5.31 HERE]

### 5.5.2.2 Assessment of Individual Components

Remaining carbon budgets are assessed through the combination of five separate components (Forster et al., 2018; Rogelj et al., 2018b). Each component is discussed and assessed separately in the sections below, based on all available lines of evidence. Box 5.1 details the differences compared to AR5 and SR1.5 estimates (Collins et al., 2013b) (Rogelj et al., 2018b).

#### 5.5.2.2.1 TCRE

The first and central component for estimating remaining carbon budgets is the TCRE. Based on the assessment in Section 5.5.1.4, an assessed *likely* range for TCRE of 1.0–2.3°C per 1000 PgC with a normal distribution is used.

#### 5.5.2.2.2 Historical Warming

Advances in methods to estimate remaining carbon budgets have shown the importance of applying an as accurate as possible estimate of historical warming to date (Millar et al., 2017a; Tokarska and Gillett, 2018). This becomes particularly important when assessing remaining carbon budgets for global warming levels that are relatively close to present-day warming, such as a 1.5°C or 2°C levels (Rogelj et al., 2018b). Also

1 the definition of global average temperature by which historical warming is estimated is shown to be  
2 important (see Cross-Chapter Box 2.3) (Cowtan and Way, 2014; Allen et al., 2018; Pflieger et al., 2018;  
3 Richardson et al., 2018; Tokarska et al., 2019a), as is the correct isolation of human-induced global warming  
4 (Haustein et al., 2017; Allen et al., 2018) to remove the effect of internal variability. Based on the assessment  
5 in Section 3.3 (Table 3.1), we here apply an assessed best-estimate historical warming estimate expressed in  
6 global average surface air temperatures (GSAT) of 1.07°C (0.8–1.3°C, *likely* range) between 1850–1900 and  
7 2010–2019. This choice implies global coverage and is consistent with AR5 where carbon budgets were  
8 reported in GSAT (Collins et al., 2013a; Stocker et al., 2013c), the SR1.5 where GSAT was the central  
9 metric for remaining carbon budgets (Rogelj et al., 2018b) and recent studies that highlight how GSAT  
10 enables an easy translation with AR5 (Tokarska et al., 2019a). The use of other historical reference periods  
11 (Cross-Chapter Box 1.2) or temperature metrics and updated data products (Cross-Chapter Box 2.3) can  
12 result in a different estimated historical warming and thus a changed remaining carbon budget.

#### 13 14 15 5.5.2.2.3 *Non-CO<sub>2</sub> Warming Contribution*

16 Non-CO<sub>2</sub> emissions contribute either cumulatively (N<sub>2</sub>O, and other long-lived climate forcers) or in  
17 proportion to their annual emissions (CH<sub>4</sub> and other short-lived climate forcers) to global warming, and thus  
18 also affect estimates of remaining carbon budgets by reducing the amount of warming that could still result  
19 from CO<sub>2</sub> emissions (Meinshausen et al., 2009; Friedlingstein et al., 2014a; Knutti and Rogelj, 2015; Rogelj  
20 et al., 2015a, 2016, Williams et al., 2017c, 2016; Matthews et al., 2017; Collins et al., 2018; Mengis et al.,  
21 2018; Tokarska et al., 2018; Zickfeld et al., 2021). The size of this contribution has been estimated both  
22 implicitly (Meinshausen et al., 2009; Friedlingstein et al., 2014a; Rogelj et al., 2016; Matthews et al., 2017;  
23 Mengis et al., 2018; Tokarska et al., 2018) and explicitly (Rogelj et al., 2015a, 2018b; Collins et al., 2018;  
24 Matthews et al., 2021) by varying the assumptions of non-CO<sub>2</sub> emissions and associated warming. Internally  
25 consistent evolutions of future CO<sub>2</sub> and non-CO<sub>2</sub> emissions allow to derive non-CO<sub>2</sub> warming contributions  
26 consistent with global CO<sub>2</sub> emissions reaching net zero levels and therewith capping maximum future CO<sub>2</sub>  
27 emissions (Smith and Mizrahi, 2013; Clarke et al., 2014; Huppmann et al., 2018; Rogelj et al., 2018b;  
28 Matthews et al., 2021). Pathways that reflect such development typically show a stabilisation or decline in  
29 non-CO<sub>2</sub> radiative forcing and warming at and after the time of global CO<sub>2</sub> emissions reaching net zero  
30 levels, as illustrated in the scenario database underlying the IPCC SR1.5 (Huppmann et al., 2018; Rogelj et  
31 al., 2018b).

32  
33 The impact of non-CO<sub>2</sub> emissions on remaining carbon budgets is assessed with emulators (Meinshausen et  
34 al., 2009; Millar et al., 2017a; Gasser et al., 2018; Goodwin et al., 2018; Rogelj et al., 2018b; Smith et al.,  
35 2018a; Matthews et al., 2021) that incorporate synthesised climate and carbon-cycle knowledge (Cross-  
36 Chapter Box 7.1). The estimated implied non-CO<sub>2</sub> warming can subsequently be applied to reduce the  
37 remaining allowable warming for estimating the remaining carbon budget (see Figure 5.31) (Rogelj et al.,  
38 2018b, 2019). Alternative methods estimate the non-CO<sub>2</sub> fraction of total anthropogenic forcing (Matthews  
39 et al., 2021), or do not correct for non-CO<sub>2</sub> warming directly. The latter methods instead consider CO<sub>2</sub> and  
40 non-CO<sub>2</sub> warming together to define a CO<sub>2</sub> forcing equivalent carbon budget from which eventual non-CO<sub>2</sub>  
41 contributions expressed in CO<sub>2</sub>-forcing-equivalent emissions have to be subtracted to obtain a remaining  
42 carbon budget (Jenkins et al., 2018; Matthews et al., 2020). These studies also use emulators to invert a  
43 specified evolution of non-CO<sub>2</sub> forcing to a corresponding amount of equivalent CO<sub>2</sub> emissions (Matthews et  
44 al., 2020), or alternatively use empirical relationships linking changes in non-CO<sub>2</sub> greenhouse gas emissions  
45 to warming (Cain et al., 2019). Methods to express non-CO<sub>2</sub> emissions in CO<sub>2</sub> equivalence are assessed in  
46 Section 7.6, yet their applicability and related uncertainties for remaining carbon budgets have not yet been  
47 covered in-depth in the literature.

48  
49 Application of the SR1.5 method (Forster et al., 2018; Rogelj et al., 2018b) with AR6-calibrated emulators  
50 (Box 7.1) suggests a median additional non-CO<sub>2</sub> warming contribution at the time global CO<sub>2</sub> emissions  
51 reach net zero levels of about 0.1–0.2°C relative to 2010–2019. Uncertainty surrounding this range due to  
52 geophysical uncertainties such as non-CO<sub>2</sub> forcing uncertainties and TCR is of the order of ±0.1°C.  
53 Differences in the choices of mitigation strategies considered in low-emission scenarios (Huppmann et al.,  
54 2018) result in a potential additional variation around the central range of at least ±0.1°C (spread across  
55 scenarios, referred to as non-CO<sub>2</sub> scenario uncertainty in Table 5.8).

#### 5.5.2.2.4 Adjustments due to the Zero-Emission Commitment (ZEC)

Use of TCRE for estimating remaining carbon budgets needs to consider the ZEC, the potential additional warming after a complete cessation of net CO<sub>2</sub> emissions. Based on the ZEC assessment presented in Section 4.7.2, the ZEC's central value is taken to be zero with a *likely* range of  $\pm 0.19^{\circ}\text{C}$ , noting that it might either increase or decrease after half a century. ZEC uncertainty is assessed for a time frame of half a century, as this most appropriately reflects the time between the time stringent mitigation pathways reach net zero CO<sub>2</sub> emissions and the end of the century. For shorter time horizons, a similar central zero value applies, but with a smaller range (MacDougall et al., 2020). Experiments that ramped up and down emissions following a bell-shaped trajectory (MacDougall and Knutti, 2016a) show that when annual CO<sub>2</sub> emissions decline to zero at a pace consistent with those currently assumed in mitigation scenarios (Huppmann et al., 2018; Rogelj et al., 2018b), the ZEC will already be realised to a large degree at the time of reaching net zero CO<sub>2</sub> emissions (MacDougall et al., 2020).

#### 5.5.2.2.5 Adjustments for Other not Represented Feedbacks

Section 5.5.1.2 highlighted recent literature describing potential impacts of Earth system feedbacks that have typically not been included in standard ESMs (Schneider von Deimling et al., 2015; MacDougall and Friedlingstein, 2015; Schädel et al., 2016; Burke et al., 2017; Mahowald et al., 2017; Comyn-Platt et al., 2018; Gasser et al., 2018; Lowe and Bernie, 2018), the most important of which is carbon release from thawing permafrost. The IPCC SR1.5 estimated unrepresented Earth system processes to result in a reduction of remaining carbon budgets of up to 100 GtCO<sub>2</sub> over the course of this century, and more thereafter (Rogelj et al., 2018b). Here this assessment is updated based on the Earth system feedback assessment of Section 5.4.8 and synthesised in Figure 5.29 by applying the reverse method by (Gregory et al., 2009).

The assessment in Section 5.4 and Box 5.1 highlights the different nature, magnitude and uncertainties surrounding additional Earth system feedback. The remaining carbon budgets reported in Table 5.8 account for these feedbacks, including corrections due to permafrost CO<sub>2</sub> and CH<sub>4</sub> feedbacks as well as those due to aerosol and atmospheric chemistry (Section 5.4.8). Two of these additional feedbacks (tropospheric ozone and methane lifetime feedbacks) are included in the projections of non-CO<sub>2</sub> warming carried out with AR6-calibrated emulators (Box 7.1). The remainder of these independent Earth system feedbacks combine to a feedback of about  $7 \pm 27 \text{ PgC K}^{-1}$  (1-sigma range, or  $26 \pm 97 \text{ GtCO}_2 \text{ }^{\circ}\text{C}^{-1}$ ). Overall, Section 5.4.8 assessed there to be *low confidence* in the exact magnitude of these feedbacks and they represent identified additional amplifying factors that scale with additional warming and mostly increase the challenge of limiting global warming to or below specific temperature levels.

#### 5.5.2.3 Remaining Carbon Budget

The combination of the five components assessed in Sections 5.5.2.2.1–5.5.2.2.5 allows for an overall assessment of the remaining carbon budget in line with different levels of global average warming, as documented in the IPCC SR1.5 (Rogelj et al., 2018b). The overall assessment of remaining carbon budgets (Table 5.8) reflects the uncertainty in TCRE quantification and provides estimates of the uncertainties surrounding the contributions of each of the respective further components. A formal combination of all uncertainties is not possible because they are not all independent or because they represent choices rather than probabilistic uncertainties (Matthews et al., 2021). In light of all uncertainties related to TCRE, non-CO<sub>2</sub> forcing and response, the level of non-CO<sub>2</sub> mitigation, and historical warming, there is a small probability that the remaining carbon budget for limiting warming to 1.5°C since the 1850–1900 period is effectively zero. However, applying best estimate values for all but uncertainties in Earth system feedbacks and TCRE, the remaining carbon budgets in line with the Paris Agreement are generally small yet not zero (see Table 5.8).

There is *robust evidence* supporting the concept of TCRE as well as *high confidence* in the range of historical human-induced warming. Combined with the assessed uncertainties in the Earth system's response

to non-CO<sub>2</sub> emissions and less well-established quantification of some of the effect of non-linear Earth system feedbacks, this leads to *medium confidence* being assigned to the assessed remaining carbon budget estimates while noting the identified and assessed uncertainties and potential variations. The reported values are applicable to warming and cumulative emissions over the 21st century. For climate stabilisation beyond the 21st century this confidence would decline to *very low confidence* due to uncertainties in Earth system feedbacks and the ZEC.

For estimates of total carbon budgets in line with limiting global warming to a specific level, an estimate of historical CO<sub>2</sub> emissions should be added to the remaining carbon budget values reported in Table 5.8. Historical CO<sub>2</sub> emissions between 1850 and 2019 have been estimated at about 655 ± 65 PgC (1-sigma range, or 2390 ± 240 GtCO<sub>2</sub>, see Table 5.1), while since 1 January 2015, an additional 57 PgC (210 GtCO<sub>2</sub>) has been emitted until the end of 2019 (Friedlingstein et al., 2020).

[START TABLE 5.8 HERE]

**Table 5.8: The assessed remaining carbon budget and corresponding uncertainties.** Assessed estimates are provided for additional human-induced warming expressed as global average surface air temperature since the recent past (2010–2019), which *likely* amounted to 0.8 to 1.3 with a best estimate of 1.07°C relative to 1850–1900 (Table 3.1 in Chapter 3).

Additional warming since 2010–2019 * <sup>(1)</sup> °C	Warming since 1850–1900 * <sup>(1)</sup> °C	Remaining carbon budget* <sup>(2)</sup> starting from 1 January 2020 and subject to variations and uncertainties quantified in the columns on the right					Scenario variation Non-CO <sub>2</sub> scenario variation * <sup>(5)</sup> PgC (GtCO <sub>2</sub> )	Geophysical uncertainties			
		Percentiles of TCRE* <sup>(3)</sup> * <sup>(4)</sup> PgC (GtCO <sub>2</sub> )						Non-CO <sub>2</sub> forcing and response uncertainty * <sup>(6)</sup> PgC (GtCO <sub>2</sub> )	Historical temperature uncertainty * <sup>(1)</sup> PgC (GtCO <sub>2</sub> )	ZEC uncertainty * <sup>(7)</sup> PgC (GtCO <sub>2</sub> )	Recent emissions uncertainty * <sup>(8)</sup> PgC (GtCO <sub>2</sub> )
		17th	33rd	50th	67th	83rd					
0.23	1.3	100 (400)	60 (250)	40 (150)	30 (100)	10 (50)	Values can vary by at least ±60 PgC (±220 GtCO <sub>2</sub> ) due to choices related to non-CO <sub>2</sub> emissions mitigation	Values can vary by at least ±60 PgC (±220 GtCO <sub>2</sub> ) due to uncertainty in the warming response to future non-CO <sub>2</sub> emissions	±150 PgC (±550 GtCO <sub>2</sub> )	±115 PgC (±420 GtCO <sub>2</sub> )	±6 PgC (±20 GtCO <sub>2</sub> )
0.33	1.4	180 (650)	120 (450)	90 (350)	70 (250)	50 (200)					
0.43	1.5	250 (900)	180 (650)	140 (500)	110 (400)	80 (300)					
0.53	1.6	330 (1200)	230 (850)	180 (650)	150 (550)	110 (400)					
0.63	1.7	400 (1450)	290 (1050)	230 (850)	190 (700)	150 (550)					
0.73	1.8	470 (1750)	350 (1250)	280 (1000)	230 (850)	180 (650)					
0.83	1.9	550 (2000)	400 (1450)	320 (1200)	270 (1000)	120 (800)					
0.93	2	620 (2300)	460 (1700)	370 (1350)	310 (1150)	250 (900)					
1.03	2.1	700 (2550)	510 (1900)	420 (1500)	360 (1250)	280 (1050)					
1.13	2.2	770 (2850)	570 (2100)	460 (1700)	390 (1400)	310 (1150)					
1.23	2.3	850 (3100)	630 (2300)	510 (1850)	430 (1550)	350 (1250)					
1.33	2.4	920 (3350)	680 (2500)	550 (2050)	470 (1700)	380 (1400)					
* <sup>(1)</sup> Human-induced global surface air temperature increase between 1850–1900 and 2010–2019 is assessed at 0.8–1.3°C ( <i>likely</i> range; Chapter 3) with a best estimate of 1.07°C. Warming here reflects GSAT, as TCRE and other estimates are GSAT based. Combined with a central estimate of TCRE (1.65 °C EgC <sup>-1</sup> ) the uncertainty in historical human-induced GSAT warming results in a potential variation of remaining carbon budgets of ±150 PgC or ±550 GtCO <sub>2</sub> .											
* <sup>(2)</sup> Historical CO <sub>2</sub> emissions between 1850 and 2019 have been estimated at about 655 ± 65 PgC (1-sigma range, or 2390 ± 240 GtCO <sub>2</sub> , see Table 5.1). Note that 57 PgC (210 GtCO <sub>2</sub> ) have been emitted from the middle of the 2010-2019 reference period (2015) until the end of 2019 (Friedlingstein et al., 2020).											
* <sup>(3)</sup> TCRE: transient climate response to cumulative emissions of carbon, assessed to fall <i>likely</i> between 1.0–2.3 °C EgC <sup>-1</sup> with a normal distribution. PgC values are rounded to the nearest 10; GtCO <sub>2</sub> values to the nearest 50. For comparison, assuming a lognormal distribution with a 1.0–2.3 °C EgC <sup>-1</sup> central 66% range instead of a normal distribution would increase remaining carbon budgets at the 17th, 33rd, 50th, 67th, and 83rd percentile with 3%, 10%, 12%, 9%, 2%, respectively. Future non-CO <sub>2</sub> contributions in these remaining carbon budget estimates are based on the scenarios assessed in the IPCC SR1.5 report and estimated as the median quantile regression of non-CO <sub>2</sub> warming since 2010–2019 relative to total additional warming since 2010–2019 at the time scenarios reach net-zero CO <sub>2</sub> emissions (Forster et al., 2018; Huppmann et al., 2018; Rogelj et al., 2018b).											
* <sup>(4)</sup> Additional Earth system feedbacks are included in the remaining carbon budget estimates as discussed in Section 5.5.2.2.5. The tropospheric ozone and methane lifetime contributions are included through the non-CO <sub>2</sub> warming projections by the AR6-calibrated MAGICC emulator, while the remaining feedbacks are assessed totalling a combined feedback of magnitude 7 ± 27 PgC K <sup>-1</sup> (1-sigma range, or 26 ± 97 GtCO <sub>2</sub> °C <sup>-1</sup> ).											
* <sup>(5)</sup> Variations due to different scenario assumptions related to the future evolution of non-CO <sub>2</sub> emissions in mitigation scenarios reaching net zero CO <sub>2</sub> emissions (Huppmann et al., 2018; Rogelj et al., 2018b) of at least ±0.1 °C (spread across scenarios). Combined with a central estimate of TCRE (1.65 °C EgC <sup>-1</sup> ) this results in at least ±60 PgC or ±220 GtCO <sub>2</sub> . This spread reflects the variation in the underlying scenario ensemble but is not a formal likelihood. WGIII will reassess the potential for non-CO <sub>2</sub> mitigation based on literature since the SR1.5.											
* <sup>(6)</sup> Remaining carbon budget variation due to geophysical uncertainty in forcing and temperature response of non-CO <sub>2</sub> emissions of the order of ±0.1 °C, very largely range (5–95%) of non-CO <sub>2</sub> response (Section 5.5.2.2.3). Combined with a central estimate of TCRE (1.65 °C EgC <sup>-1</sup> ) this results in at least ±60 PgC or ±220 GtCO <sub>2</sub> .											
* <sup>(7)</sup> The variation due to the ZEC is estimated for a central TCRE value of 1.65 °C EgC <sup>-1</sup> and a 1-sigma ZEC range of 0.19°C. In real-world pathways, the magnitude of this effect will depend on the pace of CO <sub>2</sub> emissions reductions to net zero.											

\*<sup>(8)</sup> Historical emissions uncertainty reflects the  $\pm 10\%$  uncertainty in the historical emissions estimate since 1 January 2015.

[END TABLE 5.8 HERE]

[START BOX 5.2 HERE]

**BOX 5.2: Implications of methodological advancements in estimating the remaining carbon budget since AR5**

Methodological advancements since the IPCC AR5 (Collins et al., 2013; IPCC, 2013; Clarke et al., 2014; IPCC, 2014; Stocker et al., 2013) result in an updated and strengthened assessment of remaining carbon budgets. Methods and approaches at the time of AR5 are described in Section 5.5.2.1. Since AR5, strengths and weaknesses of various approaches have been more clearly articulated in the literature (e.g., in (Rogelj et al., 2016; Millar et al., 2017a; Tokarska and Gillett, 2018; Matthews et al., 2020)), resulting in a new consolidated framework applied in SR1.5 (Rogelj et al., 2018b, 2019) that is also used in AR6. This framework incorporates five methodological advancements compared to AR5, the implications of which are discussed in this box.

First, publications since AR5 applied methods that limit the effect of uncertainties in historical, diagnosed emissions in coupled Earth system models on estimates of the remaining carbon budget (Millar et al., 2017a; Tokarska and Gillett, 2018). These new methods express remaining carbon budget estimates relative to a recent reference period instead of relative to pre-industrial (Millar et al., 2017a; Tokarska et al., 2019a). Estimates of the full carbon budget since pre-industrial can still be obtained by adding estimates of historical CO<sub>2</sub> emissions (Table 5.1) to estimates in Table 5.8. This methodological update resulted, all other aspects being equal, in median estimates of remaining carbon budgets being about 350–450 GtCO<sub>2</sub> larger compared to AR5 (IPCC, 2014; Millar et al., 2017a).

At the time of the AR5, CMIP5 (Taylor et al., 2012) provided global mean surface air temperature (GSAT) projections for the representative concentration pathways (Meinshausen et al., 2011b), which were used to determine carbon budgets while taking into account the effects of non-CO<sub>2</sub> forcings (Stocker et al., 2013c). Their use came with two recognised limitations: first, the model spread of the CMIP5 ensemble represents an ensemble of opportunity with limited statistical value (Tebaldi and Knutti, 2007); and second, the evolution of non-CO<sub>2</sub> emissions as a function of cumulative CO<sub>2</sub> emissions can differ markedly between high and low emissions pathways (Meinshausen et al., 2011a; Rogelj et al., 2016; Friedlingstein et al., 2014; Matthews et al., 2017). Solutions to these two limitations have been published since AR5 and represent the second and third methodological improvement compared to AR5.

The reliance on an ensemble of opportunity (i.e. a serendipitous collection of scenario data from a variety of sources and studies) is avoided by methodologically separating the assessment of future warming contributions of non-CO<sub>2</sub> emissions from the spread in TCRE (Rogelj et al., 2018b, 2019) (see Section 5.5.2). This facilitates the explicit representation of TCRE uncertainty by a formal distribution, in this case a normal distribution with a 1.0–2.3°C PgC<sup>-1</sup> 1-sigma range (Section 5.5.1.4). The effect of this methodological advance can be estimated from a direct comparison of the frequency distribution of TCRE in CMIP5 models that were used in AR5 and the formal TCRE distribution used in AR6, but is limited in precision. For estimates of the remaining carbon budget in line with limiting warming to 1.5°C or 2°C relative to pre-industrial levels, this improvement is estimated to lead to a reduction of budgets of the order of about 100 GtCO<sub>2</sub> between AR5 and AR6.

The third methodological improvement is a more direct estimation of the warming contribution of non-CO<sub>2</sub> emissions, consistent with pathways that bring global CO<sub>2</sub> emissions down to net zero. Instead of deriving this contribution implicitly from the CMIP5 ensemble, climate emulators (Meinshausen et al., 2011c; Schwarber et al., 2018; Smith et al., 2018b) that are calibrated to the combined AR6 assessment (Cross-Chapter Box 7.1) are used to estimate the non-CO<sub>2</sub> contribution across a wide variety of stringent mitigation



1 scenarios (Huppmann et al., 2018). The specific relative effect of this advance compared to AR5 is not  
2 calculable because CMIP5 data does not isolate non-CO<sub>2</sub> from CO<sub>2</sub>-induced warming.

3  
4 The fourth and fifth methodological advancements are to explicitly account for the zero-emission  
5 commitment (ZEC, Section 5.5.2.2.4) and adjust estimates for Earth system feedbacks that are typically not  
6 represented in Earth System models (Section 5.5.2.2.5). The central estimate of the assessed ZEC used in  
7 SR1.5 and AR6 is zero (Sections 4.7.2). ZEC uncertainties are reported separately (Table 5.8), and the  
8 additional consideration of ZEC therefore does result in a better understanding but not in a net shift of central  
9 estimates of the remaining carbon budget compared to AR5. Furthermore, AR5 did not explicitly account for  
10 Earth system feedbacks not represented in Earth system models. SR1.5 assessed that they could reduce the  
11 remaining carbon budgets by about 100 GtCO<sub>2</sub> over centennial timescales. This assessment has been updated  
12 in AR6, including a wider range of biogeochemical feedbacks and new evidence (Section 5.5.2.2.5). Some of  
13 these feedbacks are captured in the estimation of non-CO<sub>2</sub> warming (see below), while the combined effect  
14 of remaining positive and negative feedbacks is assessed to reduce the remaining carbon budget estimates by  
15  $7 \pm 27 \text{ PgC K}^{-1}$  (1-sigma range, or  $26 \pm 97 \text{ GtCO}_2 \text{ }^\circ\text{C}^{-1}$ ) compared to AR5.

16  
17 Between SR1.5 and AR6, each of the five components described in Section 5.5.2.1 and Figure 5.32 have  
18 been re-assessed (see Sections 5.5.2.2.1 to 5.5.2.2.5). Their updated assessments in turn affect the assessment  
19 of the remaining carbon budget. The new and narrower assessment of TCRE in AR6 compared to SR1.5  
20 (*likely* range of 1.0–2.3°C EgC<sup>-1</sup> compared to 0.8–2.5°C EgC<sup>-1</sup>, respectively, with the same central estimate)  
21 leads to no change in median estimates and about a 50 and 100 GtCO<sub>2</sub> increase in remaining carbon budgets  
22 estimates at the 67<sup>th</sup> percentile in AR6 compared to SR1.5 for 1.5°C and 2°C of global warming,  
23 respectively.

24  
25 For historical warming, SR1.5 used GSAT increase between 1850–1900 and 2006–2015 of 0.97°C as its  
26 main starting point, while also providing values for other temperature metrics. Remaining carbon budgets  
27 were expressed starting from 1 January 2018 by accounting for historical emissions emitted from 1 January  
28 2011 until the end of 2017. AR6 uses anthropogenic (human-induced) warming until the 2010–2019 period,  
29 which is assessed at the 0.8–1.3°C range, with a best estimate of 1.07°C (Table 3.1), and subsequently  
30 accounts for historical emissions from 1 January 2015 until the end of 2019 to express remaining carbon  
31 budget estimates from 1 January 2020 onwards. The human-induced warming between the 1850–1900 and  
32 2006–2015 periods used in SR1.5 was assessed by AR6 at 0.97°C (Table 3.1). In a like-with-like  
33 comparison, the combined effect of data and methodological updates in historical warming estimates thus  
34 results in no shift in estimated remaining carbon budgets between SR1.5 and AR6. However, the emissions  
35 of the years passed since SR1.5 reduce the remaining carbon budget by about 85 GtCO<sub>2</sub>. Note that AR6 also  
36 updated its GSAT assessment for total warming between the 1850–1900 and 2006–2015 periods, reporting  
37 0.94°C of warming. On a like-with-like basis, this would have resulted in slightly larger remaining carbon  
38 budgets compared to SR1.5 (Cross-Chapter Box 2.3).

39  
40 The non-CO<sub>2</sub> contribution to future warming in emissions scenarios (Huppmann et al., 2018) is re-assessed  
41 with AR6-calibrated emulators, in this case MAGICC7 (Meinshausen et al., 2009, 2011a, 2020) (Cross-  
42 Chapter Box 7.1). The re-assessment of non-CO<sub>2</sub> warming with MAGICC7 results in a relationship that  
43 closely matches the average relationship applied in SR1.5 (shown in Section 2.SM.1.1.2 in (Forster et al.,  
44 2018)), and does therefore not change estimates of the remaining carbon budget relative to SR1.5. The  
45 median ZEC assessment remained the same between SR1.5 and AR6, and therefore also does not change the  
46 median remaining carbon budget estimates. Finally, as indicated above, AR6 expanded the assessment of  
47 Earth system feedbacks compared to SR1.5 and included it in its central remaining carbon budget estimates.  
48 Some feedbacks are accounted for through the non-CO<sub>2</sub> warming estimate (Section 5.5.2.2.5), while the  
49 remainder combines to reduce the median remaining carbon budget estimates for 1.5°C and 2°C of warming  
50 by about 10 to 20 GtCO<sub>2</sub>, respectively, compared to SR1.5.

51  
52 All methodological improvements and new evidence combined result in median and 67<sup>th</sup> percentile  
53 remaining carbon budget estimates for limiting warming to 1.5°C being about 300–350 GtCO<sub>2</sub> larger  
54 compared to an assessment that would use the evidence and methods available at the time of the AR5. For  
55 limiting warming to 2°C, the difference is about 400–500 GtCO<sub>2</sub>. Since SR1.5, fewer key advancements had

1 to be integrated. In a like-with-like comparison, the combined effects of all AR6 updates result in median  
2 remaining carbon budget estimates for limiting warming to 1.5°C and 2°C being the same and about 60  
3 GtCO<sub>2</sub> smaller, respectively, in AR6 compared to SR1.5. At the 67<sup>th</sup> percentile, remaining carbon budget  
4 estimates for limiting warming to 1.5°C and 2°C are about 40 to 60 GtCO<sub>2</sub> larger, respectively, mainly as a  
5 result of a narrower assessed TCRE range.

6  
7 **[END BOX 5.2 HERE]**  
8  
9

## 10 **5.6 Biogeochemical Implications of Carbon Dioxide Removal and Solar Radiation Modification**

### 11 **5.6.1 Introduction**

12 Carbon dioxide removal (CDR) refers to anthropogenic activities that seek to remove CO<sub>2</sub> from the  
13 atmosphere and durably store it in geological, terrestrial or ocean reservoirs, or in products (Glossary, Annex  
14 VII). CO<sub>2</sub> is removed from the atmosphere by enhancing biological or geochemical carbon sinks or by direct  
15 capture of CO<sub>2</sub> from air and storage. Solar radiation modification (SRM), on the other hand, refers to the  
16 intentional, planetary-scale modification of the Earth's radiative budget with the aim of limiting global  
17 warming. Most proposed SRM methods involve reducing the amount of incoming solar radiation reaching  
18 the surface, but others also act on the longwave radiation budget by reducing optical thickness and cloud  
19 lifetime (Glossary, Annex VII). SRM does not fall within the IPCC definitions of mitigation and adaptation  
20 (Glossary, Annex VII). CDR and SRM are referred to as 'geoengineering' in some of the literature and are  
21 considered separately in this report.

22 This section assesses the implications of CDR and SRM for biogeochemical cycles. CDR has received  
23 growing interest as an important mitigation option in emission scenarios consistent with meeting the Paris  
24 Agreement climate goals (SR1.5, SRCCL). The climate effects of CDR and SRM are assessed in Chapter 4,  
25 and a detailed assessment of the socio-economic dimensions of these options is presented in AR6 WGIII,  
26 Chapters 7 and 12.

### 27 **5.6.2 Biogeochemical Responses to Carbon Dioxide Removal (CDR)**

28 The scope of this section is to assess the general and methods-specific effects of CDR on the global carbon  
29 cycle and other biogeochemical cycles. The focus is on Earth system feedbacks that either amplify or reduce  
30 carbon sequestration potentials of specific CDR methods, and determine their effectiveness in reducing  
31 atmospheric CO<sub>2</sub> and mitigating climate change. Technical carbon sequestration potentials of CDR methods  
32 are assessed on a qualitative scale; a comprehensive quantitative assessment is left to the AR6 Working  
33 Group III report (Chapters 7 and 12). Biogeochemical and biophysical side effects of CDR methods are  
34 assessed here while the co-benefits and trade-offs for biodiversity, water and food production are briefly  
35 discussed for completeness, but a comprehensive assessment is left to WGII (Chapters 2 and 5) and WGIII  
36 (Chapters 7 and 12). The assessment in this chapter emphasises literature published since the AR5 WGI  
37 report (Chapter 6) for the assessment of the global carbon cycle response to CDR, and literature published  
38 since the IPCC SR1.5 (Chapter 4; IPCC, 2018), SRCCL (Chapter 6, IPCC, 2019) and SROCC (Bindoff et  
39 al., 2019) for the assessment of potentials and side effects of specific CDR methods. Emerging literature on  
40 deliberate methane removal is also briefly discussed.

41 In this chapter, CDR methods are categorised by the carbon cycle processes that result in CO<sub>2</sub> removal: (i)  
42 enhanced net biological production and storage by land ecosystems, (ii) enhanced net biological production  
43 and storage in the open and coastal ocean, (iii) enhanced geochemical processes on land and in the ocean,  
44 and (iv) direct air capture and storage by chemical processes. A subset of CDR methods that restore or  
45 sustainably manage natural or modified ecosystems while providing human well-being and biodiversity  
46 benefits are also referred to as natural or nature-based solutions (see Glossary, Annex VII) (Griscom et al.,  
47 2017, 2020; Fargione et al., 2018). CDR methods commonly discussed in the literature are summarised in  
48 Table 5.9. Other CDR options have been suggested, but there is insufficient literature for an assessment.

1 These include ocean biomass burial, ocean downwelling, removal of CO<sub>2</sub> from seawater with storage, and  
 2 cloud alkalisation (Keller et al., 2018a; GESAMP, 2019).  
 3  
 4

5 **[START TABLE 5.9 HERE]**

6  
 7 **Table 5.9: Characteristics of carbon dioxide removal (CDR) methods.** Termination effects refer to the possible  
 8 effects of a hypothetical, sudden and sustained termination of the CDR method.  
 9

Category	Methods (subsection where the method is assessed)	Nature of CO <sub>2</sub> Removal Process / Storage Form	Description	Time scale of carbon storage	Factors that affect carbon storage time scale	Termination effects
Enhanced biological production and storage on land (in vegetation, soils or geologic formations)	Afforestation, reforestation and forest management (5.6.2.2.1)	Biological / Organic	Store carbon in trees and soils by planting, restoring or managing forests	Decades to centuries (Cooper, 1983)	Disturbances (e.g., fires, pests), extreme weather	None
	Soil carbon sequestration (5.6.2.2.1)	Biological / Organic	Use agricultural management practices to improve soil carbon storage	Decades to centuries (Dignac et al., 2017)	Soil and crop management	None
	Biochar (5.6.2.2.1)	Biological / Organic	Burn biomass at high temperature under anoxic conditions to form biochar and add to soils	Decades to centuries (Campbell et al., 2018)	Fire	None
	Peatland restoration (5.6.2.2.1)	Biological / Organic	Store carbon in soil by creating or restoring peatlands	Decades to centuries (Harenda et al., 2018)	Peatland drainage, fire, drought, land use change	None
	Bioenergy with carbon capture and storage (BECCS) (5.6.2.2.1)	Biological / Inorganic	Production of energy from plant biomass combined with carbon capture and storage	Potentially permanent (analogous to DACCS) (Szulczewski et al., 2012)	Leakage	None
Enhanced biological production and storage in coastal and open ocean	Ocean fertilisation (5.6.2.2.2)	Biological / organic	Fertilise upper ocean with micro (Fe) and macronutrients (N, P) to increase phytoplankton photosynthesis and biomass and deep ocean carbon storage through the biological	Decades to millennia (Oschlies et al., 2010; Robinson et al., 2014)	Ocean stratification and circulation (Robinson et al., 2014); efficiency of carbon sequestration in deep ocean (Yoon et al., 2018)	Uncertain (Keller et al., 2014)

	Artificial ocean upwelling (5.6.2.2.2)	Biological / organic	pump Pump nutrient-rich deep ocean water to the surface to increase carbon uptake and storage through the biological pump.	Centuries to millennia (Oschlies et al., 2010b)	Ocean circulation; DIC content of upwelled waters (Oschlies et al., 2010c)	Warming beyond temperatures experienced if artificial ocean upwelling had not been deployed (Keller et al., 2014)
	Restoration of vegetated coastal ecosystems (“blue carbon”) (5.6.2.2.2)	Biological / organic	Manage coastal ecosystems to increase net primary production and store carbon in sediments	Decades to centuries if functional integrity of ecosystem maintained (McLeod et al., 2011)	Land use change of coastal ecosystems; extreme weather (e.g., heatwaves); sea level change (NASSEM, 2019)	None
Enhanced geochemical processes on land and in ocean	Enhanced weathering (5.6.2.2.3)	Geochemical / inorganic	Spread alkaline minerals on land to chemically remove atmospheric CO <sub>2</sub> in reactions that form solid minerals (carbonates and silicates) that are stored in soils or in the ocean	10,000 to 10 <sup>6</sup> years (Fuss et al., 2018)	Storage in soils or ocean (Fuss et al., 2018)	None
	Ocean alkalisation (5.6.2.2.3)	Geochemical / inorganic	Increased CO <sub>2</sub> uptake via increased alkalinity by deposition of alkaline minerals (e.g. olivine).	10,000 to 100,000 years (Keller, 2019)	Carbonate chemistry; ocean stratification and circulation (Keller, 2019)	Higher rates of warming and acidification than if alkalisation not begun (under a high emissions scenario) (González et al., 2018)
Chemical	Direct air carbon capture with storage (DACCS) (5.6.2.2.4)	Chemical / inorganic	Direct removal of CO <sub>2</sub> from air through chemical adsorption, absorption or mineralisation, and storage underground,	Potentially permanent	Leakage	None

			in deep ocean or in long- lasting usable materials			
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1  
2 **[END TABLE 5.9 HERE]**

3  
4  
5 **5.6.2.1 Global Carbon Cycle Responses to CDR**

6  
7 This subsection assesses evidence about the response of the global carbon cycle to CDR from idealised  
8 model simulations which assume that CO<sub>2</sub> is removed from the atmosphere directly and stored permanently  
9 in the geologic reservoir, which is analogous to direct air carbon capture with carbon storage (DACCS;  
10 Table 5.9). The carbon cycle response to specific land and ocean-based CDR methods is assessed in Section  
11 5.6.2.2.2. At the time of AR5 there were very few studies about the global carbon cycle response to CDR.  
12 Based on these studies and general understanding of the carbon cycle, AR5 WGI Chapter 6 assessed that it is  
13 *virtually certain* that deliberate removal of CO<sub>2</sub> from the atmosphere will be partially offset by outgassing of  
14 CO<sub>2</sub> from the ocean and land carbon sinks. *Low confidence* was placed on any quantification of effects.  
15 Since the WGI AR5 (Chapter 6), several studies have investigated the carbon cycle response to CDR in  
16 idealised “pulse” removal simulations, whereby a specified amount of CO<sub>2</sub> is removed instantly from the  
17 atmosphere, and scenario simulations with CO<sub>2</sub> emissions and removals following a plausible trajectory. In  
18 addition, a dedicated carbon dioxide removal model intercomparison project (CDRMIP) (Keller et al.,  
19 2018b) was initiated, which includes a range of CDR experiments from idealised simulations to simulations  
20 of deployment of specific CDR methods (afforestation and ocean alkalisation).

21  
22 This subsection assesses three aspects of the climate-carbon cycle response to CDR: the time-dependent  
23 behaviour of CO<sub>2</sub> fluxes in scenarios with CDR, the effectiveness of CDR in drawing down atmospheric CO<sub>2</sub>  
24 and cooling global mean temperature, and the symmetry of the climate-carbon cycle response to positive and  
25 negative CO<sub>2</sub> emissions.

26  
27  
28 **[START BOX 5.3 HERE]**

29  
30 **BOX 5.3: Carbon cycle response to CO<sub>2</sub> removal from the atmosphere**

31  
32 During the industrial era, CO<sub>2</sub> emitted by the combustion of fossil fuels and land-use change has been  
33 redistributed between atmosphere, land, and ocean carbon reservoirs due to carbon cycle processes (Box 5.3  
34 Figure 1b; Figure 5.13). Over the past decade (2010–2019), 46% of the emitted CO<sub>2</sub> remained in the  
35 atmosphere, 23% was taken up by the ocean and 31% by the terrestrial biosphere (Section 5.2.1.5). When  
36 carbon dioxide removal (CDR) is applied during periods in which human activities are net CO<sub>2</sub> sources to  
37 the atmosphere and the amount of emissions removed by CDR is smaller than the net source (net positive  
38 CO<sub>2</sub> emissions), CDR acts to reduce the net emissions (Box 5.3 Figure 1c). In this scenario part of the CO<sub>2</sub>  
39 emissions into the atmosphere is removed by the land and ocean sinks, as has been the case historically.

40  
41 When CDR removes more CO<sub>2</sub> emissions than human activities emit (net negative CO<sub>2</sub> emissions), and  
42 atmospheric CO<sub>2</sub> declines, the land and ocean sinks initially continue to take up CO<sub>2</sub> from the atmosphere.  
43 This because carbon sinks, particularly the ocean, exhibit inertia and continue to respond to the prior  
44 trajectory of rising atmospheric CO<sub>2</sub> concentration. After some time, which is determined by the magnitude  
45 of the removal and the rate and amount of CO<sub>2</sub> emissions prior to the CDR application, land and ocean  
46 carbon reservoirs begin to release CO<sub>2</sub> to the atmosphere making CDR less effective (Box 5.3 Figure 1d).

47  
48  
49 **[START BOX 5.3, FIGURE 1 HERE]**

50  
51 **Box 5.3, Figure 1: Schematic representation of carbon fluxes between atmosphere, land, ocean and geological**

**reservoirs.** Different system conditions are shown: (a) an unperturbed Earth system; and changes in carbon fluxes for (b) an Earth system perturbed by fossil-fuel CO<sub>2</sub> emissions, (c) an Earth system in which fossil-fuel CO<sub>2</sub> emissions are partially offset by CDR, (d) an Earth system in which CDR exceeds CO<sub>2</sub> emissions from fossil fuels (“net negative” CO<sub>2</sub> emissions). Carbon fluxes depicted in (a) (solid and dashed black lines) also occur in (b)-(d). The question mark in the land-to-ocean carbon flux perturbation in (c) and (d) indicates that the effect of CDR on this flux is unknown. Note that box sizes do not scale with the size of carbon reservoirs. Adapted from (Keller et al., 2018a). Further details on data sources and processing are available in the chapter data table (Table 5.SM.6).

[END BOX 5.3, FIGURE 1 HERE]

[END BOX 5.3 HERE]

#### 5.6.2.1.1 Carbon Cycle Response to Instantaneous CDR

Idealised “pulse” removal Earth system model simulations are useful for understanding the carbon cycle response to CDR. Figure 5.32 illustrates the response of atmospheric CO<sub>2</sub>, land and ocean carbon sinks to an instantaneous CO<sub>2</sub> removal applied from a pre-industrial equilibrium state. Following CO<sub>2</sub> removal from the atmosphere, the atmospheric CO<sub>2</sub> concentration declines rapidly at first and then rebounds (Figure 5.32 a). This rebound is due to CO<sub>2</sub> release by the terrestrial biosphere and the ocean in response to declining atmospheric CO<sub>2</sub> levels (Figure 5.32 b, c) (Collins et al., 2013a). For the model simulations shown in Figure 5.32,  $23 \pm 6\%$  (mean  $\pm 1$  standard deviation) of the 100 PgC removed remains out of the atmosphere 80–100 years after the instantaneous removal. The remainder is offset by CO<sub>2</sub> outgassing from the land ( $49 \pm 12\%$ ) and ocean ( $29 \pm 7\%$ ). While the direction of the CO<sub>2</sub> flux is robust across models, the relative contribution of the outgassing from land and ocean reservoirs to the atmospheric CO<sub>2</sub> rebound after removal varies. These results corroborate the *high confidence* placed by WGI AR5 Chapter 6 on the partial compensation of CO<sub>2</sub> removal from the atmosphere by CO<sub>2</sub> outgassing from the land and ocean. Due to disagreement between models, the magnitude of this outgassing and in the relative contribution of land and ocean fluxes remains *low confidence*.

[START FIGURE 5.32 HERE]

#### Figure 5.32: Carbon cycle response to instantaneous carbon dioxide (CO<sub>2</sub>) removal from the atmosphere. (a)

Atmospheric CO<sub>2</sub> concentration, (b) change in land carbon reservoir, (c) change in ocean carbon reservoir. Results are shown for simulations with seven CMIP6 Earth system models and the UVic ESCM model of intermediate complexity forced with 100 PgC instantaneously removed from the atmosphere. The ‘pulse’ removal is applied from a model state in equilibrium with a pre-industrial atmospheric CO<sub>2</sub> concentration (CDRMIP experiment CDR-pi-pulse; Keller et al., 2018b). Changes in land and ocean carbon reservoirs are calculated relative to a pre-industrial control simulation. Data for the UVic ESCM is from Zickfeld et al. (2021). Further details on data sources and processing are available in the chapter data table (Table 5.SM.6).

[END FIGURE 5.32 HERE]

#### 5.6.2.1.2 Carbon Cycle Response Over Time in Scenarios with CDR

Since WGI AR5 (Chapter 6), studies with ESMs have explored the land and ocean carbon sink response to scenarios with CO<sub>2</sub> emissions gradually declining during the 21st century. As CDR and other mitigation activities are ramped up, CO<sub>2</sub> emissions in these scenarios reach net zero and, as removals exceed emissions, become net negative. Studies exploring the carbon sink response to such scenarios (e.g. RCP2.6, SSP1–2.6) show that when net CO<sub>2</sub> emissions are positive, but start to decline, uptake of CO<sub>2</sub> by the land and ocean begins to weaken (compare land and ocean CO<sub>2</sub> fluxes in panels (a) and (b) of Figure 5.33) (Tokarska and Zickfeld, 2015; Jones et al., 2016b). During the first decades after CO<sub>2</sub> emissions become net negative, both the ocean and land carbon sinks continue to take up CO<sub>2</sub>, albeit at a lower rate. For the land carbon sink, the sink-to-source transition occurs decades to a century after CO<sub>2</sub> emissions become net negative (Figure

5.33c). The ocean remains a sink of CO<sub>2</sub> for centuries after emissions become net negative (Figure 5.33c-e; Section 5.4.9; Figure 5.30). Whether the transition to source occurs at all, the timing of the transition and the magnitude of the CO<sub>2</sub> source are determined by the magnitude of the removal and the rate and amount of net CO<sub>2</sub> emissions prior to emissions becoming net negative (*medium confidence*) (Tokarska and Zickfeld, 2015; Jones et al., 2016b). For scenarios with large amounts of CO<sub>2</sub> removal such as SSP5–3.4–overshoot the land source is larger than for SSP1–2.6 and the ocean also turns into a source (Section 5.4.10, Figure 5.30). While the qualitative response to scenarios with net-negative emissions is largely robust across models, the timing of the sink-to-source transition and the magnitude of the CO<sub>2</sub> source vary between models, particularly for the land sink. Due to *limited agreement* between models there is *low confidence* in the timing of the sink-to-source transition and the magnitude of the CO<sub>2</sub> source in scenarios with net-negative CO<sub>2</sub> emissions.

[START FIGURE 5.33 HERE]

**Figure 5.33: Carbon sink response in a scenario with net carbon dioxide (CO<sub>2</sub>) removal from the atmosphere.**

Shown are CO<sub>2</sub> flux components from concentration-driven Earth system model simulations during different emission stages of SSP1–2.6 and its long-term extension. (a) Large net positive CO<sub>2</sub> emissions, (b) small net positive CO<sub>2</sub> emissions, (c) – (d) net negative CO<sub>2</sub> emissions, (e) net zero CO<sub>2</sub> emissions. Positive flux components act to raise the atmospheric CO<sub>2</sub> concentration, whereas negative components act to lower the CO<sub>2</sub> concentration. Net CO<sub>2</sub> emissions, land and ocean CO<sub>2</sub> fluxes represent the multi-model mean and standard deviation (error bar) of four ESMs (CanESM5, UKESM1, CESM2-WACCM, IPSL-CM6a-LR) and one EMIC (UVic ESCM; (Mengis et al., 2020)). Net CO<sub>2</sub> emissions are calculated from concentration-driven Earth system model simulations as the residual from the rate of increase in atmospheric CO<sub>2</sub> and land and ocean CO<sub>2</sub> fluxes. Fluxes are accumulated over each 50-year period and converted to concentration units (ppm). Further details on data sources and processing are available in the chapter data table (Table 5.SM.6).

[END FIGURE 5.33 HERE]

#### 5.6.2.1.3 Removal Effectiveness of CDR

It is well understood that land and ocean carbon fluxes are sensitive to the level of atmospheric CO<sub>2</sub> and climate change and differ under different future scenarios (Section 5.4). It is therefore important to establish to what extent the removal effectiveness of CDR – here defined as the fraction of total CO<sub>2</sub> removed remaining out of the atmosphere – is dependent on the scenario from which CDR is applied. Different metrics have been proposed to quantify the removal effectiveness of CDR (Tokarska and Zickfeld, 2015; Jones et al., 2016b; Zickfeld et al., 2016). One is the airborne fraction of cumulative CO<sub>2</sub> emissions (AF), defined in the same way as for positive emissions (i.e. as the fraction of total CO<sub>2</sub> emissions remaining in the atmosphere), with its use extended to periods of declining and net negative CO<sub>2</sub> emissions. This metric, however, has not proven to be useful to quantify the removal effectiveness of CDR in simulations where CDR is applied from a trajectory of increasing atmospheric CO<sub>2</sub> concentration, as it measures the carbon cycle response to CDR as well as to the prior atmospheric CO<sub>2</sub> trajectory (Tokarska and Zickfeld, 2015; Jones et al., 2016b). A more useful metric is the perturbation airborne fraction (PAF) (Jones et al., 2016b), which measures the airborne fraction of the perturbation (in this case the CO<sub>2</sub> removal) relative to a reference scenario (Tokarska and Zickfeld, 2015; Jones et al., 2016b). The advantage of this metric is that it isolates the response to a CO<sub>2</sub> removal from the response to atmospheric CO<sub>2</sub> prior to the point in time the removal is applied. A disadvantage is that the PAF cannot be calculated from a single model simulation but requires a reference simulation relative to which the effect of the CO<sub>2</sub> removal can be evaluated. When CDR is applied from an equilibrium state, the PAF and AF are equivalent measures.

In scenario simulations and idealised simulations with instantaneous CO<sub>2</sub> removals applied from an equilibrium state, the removal effectiveness of CDR is found to be slightly dependent on the rate and amount of CDR (Tokarska and Zickfeld, 2015; Jones et al., 2016b; Zickfeld et al., 2021), and to be strongly dependent on the emission scenario from which CDR is applied (Jones et al., 2016b; Zickfeld et al., 2021). The fraction of CO<sub>2</sub> removed remaining out of the atmosphere decreases slightly for larger removals and decreases strongly when CDR is applied from a lower background atmospheric CO<sub>2</sub> concentration (Figure

5.34), due to state dependencies and climate-carbon cycle feedbacks that lead to a stronger overall response to CO<sub>2</sub> removal (Zickfeld et al., 2021). Based on the *high agreement* between studies we assess with *medium confidence* that the removal effectiveness of CDR is only slightly dependent on the rate and magnitude of removal and is smaller at lower background atmospheric CO<sub>2</sub> concentrations. Simulations with Earth system models of EMIC with instantaneous CO<sub>2</sub> removal from different equilibrium initial states suggest that the smaller removal effectiveness of CDR at lower background CO<sub>2</sub> levels results in greater cooling per unit CO<sub>2</sub> removed (Zickfeld et al., 2021). However, there is *low confidence* in the robustness of this result as climate sensitivity has been shown to exhibit opposite state dependence in EMICs and ESMs (Section 7.4.3.1).

[START FIGURE 5.34 HERE]

**Figure 5.34: Removal effectiveness of carbon dioxide removal (CDR).** (a) Fraction of CO<sub>2</sub> remaining out of the atmosphere for idealised model simulations with CDR applied instantly (pulse removals) from climate states in equilibrium with different atmospheric CO<sub>2</sub> concentration levels (1 to 4 times the pre-industrial atmospheric CO<sub>2</sub> concentration; shown on the horizontal axis). The fraction is calculated 100 years after pulse removal. The black triangle and error bar indicate the multi-model mean and standard deviation for the seven Earth system models shown in Figure 5.32 forced with a 100 PgC pulse removal. Other symbols illustrate results with the UVic ESCM model of intermediate complexity for different magnitudes of pulse removals (triangles: -100 PgC; circles: -500 PgC; squares: -1000 PgC). Data for the UVic ESCM is from (Zickfeld et al., 2021). (b) Perturbation airborne fraction (see text for definition) for model simulations where CDR is applied from four RCPs (shown on the horizontal axis in terms of their cumulative CO<sub>2</sub> emissions during 2020–2099). Symbols indicate results for four CDR scenarios, which differ in terms of the magnitude and rate of CDR (see Jones et al. (2016b) for details). Results are based on simulations with the Hadley Centre Simple Climate-Carbon Model and are shown for the year 2100. Data from Jones et al. (2016b). Further details on data sources and processing are available in the chapter data table (Table 5.SM.6).

[END FIGURE 5.34 HERE]

#### 5.6.2.1.4 Symmetry of Carbon Cycle Response to Positive and Negative CO<sub>2</sub> Emissions

It is commonly assumed that the climate-carbon cycle response to a negative CO<sub>2</sub> emission (i.e. removal from the atmosphere) is equal in magnitude and opposite in sign to the response to a positive CO<sub>2</sub> emission of equal magnitude, that is, symmetric. If the response were symmetric, a positive CO<sub>2</sub> emission could be offset by a negative emission. This subsection assesses the symmetry in the coupled climate-carbon cycle response in model simulations with positive and negative CO<sub>2</sub> emission pulses applied from a pre-industrial climate state. Simulations with seven CMIP6 ESMs and the UVic ESCM model of intermediate complexity suggest that the carbon cycle response is asymmetric for pulse emissions/removals of  $\pm 100$  PgC (Figure 5.35). For all models, the fraction of CO<sub>2</sub> remaining in the atmosphere after an emission is larger than the fraction of CO<sub>2</sub> remaining out of the atmosphere after a removal (by  $4 \pm 3\%$ ; mean  $\pm$  standard deviation). In other words, an emission of CO<sub>2</sub> into the atmosphere is more effective at raising atmospheric CO<sub>2</sub> than an equivalent CO<sub>2</sub> removal is at lowering it. Sensitivity experiments with the UVic ESCM suggest that the asymmetry increases for larger amounts of emissions/removals and is insensitive to the background atmospheric CO<sub>2</sub> concentration from which the emissions/removals are applied (Figure 5.35). This asymmetry in the atmospheric CO<sub>2</sub> response originates from asymmetries in the land and ocean carbon fluxes due to nonlinearities in the carbon cycle response to CO<sub>2</sub> and temperature (Section 5.4) (Zickfeld et al., 2021). Given *medium evidence* and *high agreement*, there is *medium confidence* in the sign of the asymmetry of the carbon cycle response to positive and negative CO<sub>2</sub> emissions. The sign of the symmetry of the temperature response, on the other hand, differs between models, with three out of seven examined ESMs showing a smaller temperature response to a 100 PgC CO<sub>2</sub> emission than to an equivalent CO<sub>2</sub> removal. Therefore, there is *low confidence* in the sign of the asymmetry of the temperature response to positive and negative CO<sub>2</sub> emissions.



1 [START FIGURE 5.35 HERE]

2  
3 **Figure 5.35: Asymmetry in the atmospheric carbon dioxide (CO<sub>2</sub>) response to CO<sub>2</sub> emissions and removals.**

4 Shown are the fractions of total CO<sub>2</sub> emissions remaining in the atmosphere (right-hand side) and CO<sub>2</sub>  
5 removals remaining out of the atmosphere (left-hand side) 80–100 after a pulse emission/removal.  
6 Triangles and green circles denote results for seven Earth system models (ESMs) and the UVic ESCM  
7 model of intermediate complexity forced with ±100 PgC pulses applied from a pre-industrial state  
8 (1×CO<sub>2</sub>) (CDRMIP experiment CDR-pi-pulse; Keller et al. (2018b)). Yellow circles and diamonds  
9 indicate UVic ESCM results for CO<sub>2</sub> emissions/removals applied at 1.5 times (1.5×CO<sub>2</sub>) and 2 times  
10 (2×CO<sub>2</sub>) the pre-industrial CO<sub>2</sub> concentration, respectively. Pulses applied from a 2×CO<sub>2</sub> state span the  
11 magnitude ±100 PgC to ±500 PgC. UVic ESCM data is from (Zickfeld et al., 2021). Further details on  
12 data sources and processing are available in the chapter data table (Table 5.SM.6).

13  
14 [END FIGURE 5.35 HERE]

15  
16  
17 *5.6.2.2 Effects of Specific CDR Methods on Biogeochemical Cycles and Climate*

18  
19 WGI AR5 Chapter 6 discussed the CDR methods, their implications and unintended side effects on carbon  
20 cycle and climate, including their time scales and potentials. Since then, three IPCC special reports (SR)  
21 have been published. First, SR15 Chapter 4 (IPCC, 2018a) assessed the potentials and current understanding,  
22 including the side effects, of BECCS, afforestation/reforestation, soil carbon sequestration, biochar,  
23 enhanced weathering, ocean alkalisation, DACCS and ocean fertilisation. Second, SRCCL Chapter 6  
24 (IPCC, 2019a) assessed the potentials, co-benefits and trade-offs of land-based mitigation options. It  
25 assessed with *high confidence* that land-based CDR options do not sequester carbon indefinitely, except for  
26 peatland restoration. Multiple co-benefits were identified in the deployment of CDR options, many of them  
27 with a potential to make positive contributions to sustainable development, enhancement of ecosystem  
28 functions and services and other societal goals. However, their potential was concluded to be context specific  
29 and limits to their contribution to global mitigation, such as competition for land, were identified. The third  
30 report, SROCC Chapter 5 (IPCC, 2019b), assessed the potential of marine options for climate change  
31 mitigation. It concluded that the feasibility of open ocean fertilisation and alkalisation approaches were  
32 negligible, due to their inconclusive influence on ocean carbon storage on long timescales, due to the  
33 unintended side effects on marine ecosystems, and the associated governance challenges. The assessment of  
34 the benefits of blue carbon ecosystems concluded that they could contribute only minimally to atmospheric  
35 CO<sub>2</sub> reduction globally but emphasised that the benefits of protection and restoration of coastal blue carbon  
36 extend beyond climate mitigation (SROCC Section 5.5.12).

37  
38  
39 *5.6.2.2.1 Land-based Biological CDR Methods*

40 Biological CDR methods, introduced in Table 5.9, seek to increase carbon storage on land by enhancing net  
41 primary productivity and/or reducing CO<sub>2</sub> sources to the atmosphere.

42  
43 Forest-based methods include afforestation, reforestation, and forest management (Table 5.9). Building on  
44 previous work that emphasized the global potentials of various options, more recent advances have focused  
45 on the limits of those global potentials in light of ecological and climate risks that can threaten the long-term  
46 permanence of carbon stocks (Boysen et al., 2017b; Anderegg et al., 2020). Some of those risks arise from  
47 droughts, fires, insect outbreaks, diseases, erosion, and other disturbances (Thompson et al., 2009).

48  
49 Sustainable forest management can help to manage some of these vulnerabilities, while in some cases, it can  
50 increase and maintain forest sinks through harvest, transfer of carbon to wood products and their use to store  
51 carbon and substitute emissions-intensive construction materials (Churkina et al., 2020). Forest genomics  
52 techniques can increase the success of both reforestation and conservation initiatives, accelerating breeding  
53 for tree health and productivity (Isabel et al., 2020).

54  
55 In response to increasing risks to permanence of carbon stocks of some types of afforestation practices and  
56 the competition for land, there has been an increasing recognition that secondary forest regrowth and

1 restoration of degraded forests and non-forest ecosystems can play a large role in carbon sequestration (*high*  
2 *confidence*). The rationale for this focus builds on their high carbon stocks and rates of sequestration  
3 (Griscom et al., 2017; Lewis et al., 2019; Maxwell et al., 2019; Pugh et al., 2019b), high resilience to  
4 disturbances (Dymond et al., 2014; Messier et al., 2019), and more additional benefits such as enhanced  
5 biodiversity (Strassburg et al., 2020).

6  
7 The global sequestration potential of forestation varies substantially depending on the scenario-assumptions  
8 of available land and of background climate (WGIII AR6 Section 7.5). Afforestation of native grasslands,  
9 savannas, and open-canopy woodlands leads to the undesirable loss of unique natural ecosystems with rich  
10 biodiversity, carbon storage and other ecosystem services (Veldman et al., 2015; IBPES, 2018).  
11 Comprehensive approaches to assess the effectiveness of land-based carbon removal options need to be  
12 based on the whole carbon cycle covering both carbon stocks and flows, and establishing the links between  
13 human activities and their impacts on the biosphere and atmosphere (Keith et al., 2021).

14  
15 A range of mechanisms could enhance CO<sub>2</sub> sequestration of forest-based methods under future scenarios,  
16 including CO<sub>2</sub> fertilisation, soil carbon enrichment due to enhanced litter input, or the northward shift of the  
17 tree-line in future climate projection (Bathiany et al., 2010; Sonntag et al., 2015; Boysen et al., 2017; Harper  
18 et al., 2018). There is *low confidence* in the net direction of feedbacks of afforestation on global mean  
19 temperature. The feedbacks are highly region dependent. For instances, afforestation at high latitudes would  
20 decrease albedo and increase local warming, while at low latitudes, the cooling effect of enhanced  
21 evapotranspiration could exceed the warming effect due to albedo decrease (Pearson et al. 2013; Zhang et al.  
22 2013; Jia et al. 2019, SRCCL Section 2.6.1). Both afforestation and reforestation affect the hydrological  
23 cycle through increased VOC emissions and cloud albedo (Teuling et al., 2017; Kalliokoski et al., 2020),  
24 enhanced precipitation (Ellison et al., 2017) and increased transpiration, with potential effects on runoff and,  
25 especially in dry areas, on water supply (Farley et al., 2005; Smith et al., 2016; Krause et al., 2017; Teuling  
26 et al., 2019) (Figure 5.36, Cross-Chapter Box 5.1). Forest-based methods can either raise or lower N<sub>2</sub>O  
27 emissions, depending on tree species, previous land use, soil type and climatic factors (*low confidence*)  
28 (Benanti et al., 2014; Chen et al., 2019; McDaniel et al., 2019) (Figure 5.36, Supplementary Materials, Table  
29 5.SM.4). Afforestation will decrease biodiversity if native species are replaced by monocultures (*high*  
30 *confidence*), while there is *medium confidence* that biodiversity is improved when forests are introduced into  
31 land areas with degraded soils or intensive monocultures, or where native species are re-introduced into  
32 managed land (Hua et al., 2016; Williamson, P., & Bodle, 2016; Smith et al., 2018c; Holl and Brancalion,  
33 2020) (Figure 5.36, Supplementary Materials Table 5.SM.4).

34  
35 Soil carbon losses from human agriculture accounted for about 116 PgC in the last 12,000 years (Sanderman  
36 et al., 2017), (Section 5.2.1.2). With best management practices, two-thirds of these losses may be  
37 recoverable, setting a theoretical maximum of 77 PgC that can be sequestered in soils. Methods to increase  
38 soil carbon content may be applied to the restoration of marginal or degraded land (Paustian et al., 2016;  
39 Smith, 2016), but may also be used in traditional agricultural lands. A simple practice is to increase the input  
40 of carbon to the soil by selecting appropriate varieties or species with greater root mass (Kell, 2011) or  
41 higher yields and NPP (Burney et al., 2010). In addition, improved agricultural practices also increase soil  
42 carbon content. These include the use of crop rotation cycles, increase the amount of crop residues, use of  
43 crop cover to prevent periods of bare soil (Poeplau and Don, 2015; Griscom et al., 2017), optimisation of  
44 grazing (Henderson et al., 2015) and residue management (Wilhelm et al., 2004), use of irrigation (Campos  
45 et al., 2020), employment of low-tillage or no-tillage (Sun et al., 2020b), agroforestry, and cropland nutrient  
46 recycling, and avoid grassland conversion management (Paustian et al., 2016; Fargione et al., 2018). With  
47 *medium confidence*, methods which seek soil carbon sequestration will diminish N<sub>2</sub>O emissions and nutrient  
48 leaching, and improve soil fertility and biological activity (Tonitto et al., 2006; Fornara et al., 2011; Paustian  
49 et al., 2016a; Smith et al., 2016; SRCCL 2.6.1.3, IPCC, 2019a; Figure 5.36). However, if improved soil  
50 carbon sequestration practices involve higher fertilisation rates, N<sub>2</sub>O emissions would increase (Gu et al.,  
51 2017). Some soil carbon sequestration methods, such as cover crops and crop diversity, can increase  
52 biodiversity (*medium confidence*) (Paustian et al., 2016; Smith et al., 2018).

53  
54 Biochar is produced by burning biomass at high temperatures under anoxic conditions (pyrolysis) and can,  
55 when added to soils, increase soil carbon stocks and fertility for decades to centuries (Woolf et al., 2010;

1 Lehmann et al., 2015). Biochar application improves many soil qualities and increase crop yield (*medium*  
2 *confidence*) (Ye et al., 2020; SRCCL Chapter 4.9.5), particularly in already degraded or weathered soils  
3 (Woolf et al., 2010; Lorenz and Lal, 2014; Jeffery et al., 2016), increases soil water holding capacity  
4 (*medium confidence*) (Karhu et al., 2011; Liu et al., 2016; Fischer et al., 2019a; Verheijen et al., 2019) and  
5 evapotranspiration (*low confidence*) (Fischer et al., 2019). The use of biochar reduces nutrient losses (*low*  
6 *confidence*) (Woolf et al., 2010), enhances fertiliser nitrogen use efficiency and improves the bioavailability  
7 of phosphorus (Clough et al., 2013; Shen et al., 2016; Liu et al., 2017b; Figure 5.36). Biochar addition may  
8 decrease CH<sub>4</sub> emissions in inundated and acid soils such as rice fields (*low confidence*) (Jeffery et al., 2016;  
9 Huang et al., 2019; Wang et al., 2019a; Yang et al., 2019a). In non-inundated, neutral soils CH<sub>4</sub> uptake from  
10 the atmosphere is suppressed after biochar application (*low confidence*) (Jeffery et al., 2016), and soil N<sub>2</sub>O  
11 emissions decline (*medium confidence*) (Cayuela et al., 2014; Kammann et al., 2017). Potential risks of  
12 introducing harmful contaminants into the soil environment are not well understood (Lorenz and Lal, 2014).  
13 With *low confidence*, application of biochar can have co-benefits for soil microbial biodiversity (Smith et al.,  
14 2018), while the potential trade-offs for biodiversity are due to land requirements (Tisserant and Cherubini,  
15 2019).

16  
17 Peatlands are less extensive than forests, croplands and grazing lands, yet per unit area, they hold high  
18 carbon stocks (Griscom et al., 2017). Peatland restoration relies on back-conversion or building of high-  
19 carbon-density soils through flooding, that is rewetting (Leifeld et al., 2019). High water level and anoxic  
20 conditions are prerequisites for restoring by returning drained and/or degraded peatlands back to their natural  
21 state as CO<sub>2</sub> sinks, but restoration also results in enhanced CH<sub>4</sub> emissions which are similar or higher than  
22 the pre-drainage fluxes (*high confidence*) (Koskinen et al., 2016; Wilson et al., 2016a; Hemes et al., 2019;  
23 Renou-Wilson et al., 2019; Holl et al., 2020). In a multi-decadal timeframe, the reduction in CO<sub>2</sub> emissions  
24 from rewetting more than compensates for the initial increase in radiative forcing due to enhanced CH<sub>4</sub>  
25 emissions (Günther et al., 2020). Rewetting drained peatlands will decrease N<sub>2</sub>O emissions (*medium*  
26 *confidence*) (Wilson et al., 2016b; Liu et al., 2020a; Tiemeyer et al., 2020). Restored wetlands and peatlands  
27 act as buffer zones that provide infiltration and nutrient retention and offer protection to water quality  
28 (Daneshvar et al., 2017; Lundin et al., 2017), particularly in nutrient-loaded agricultural catchments.  
29 Peatland restoration can also recover much of the original biodiversity (*medium confidence*) (Meli et al.,  
30 2014; Smith et al. 2018).

31  
32 The concept of bioenergy with carbon capture and storage (BECCS) rests on the premise that bioenergy  
33 production is carbon neutral, that is as much CO<sub>2</sub> is sequestered when growing biomass as feedstock as is  
34 released by its combustion. If these emissions are also captured and stored, the net effect is removal of CO<sub>2</sub>  
35 from the atmosphere (Fuss et al., 2018). Sequestration potentials from BECCS depend strongly on the  
36 feedstock, climate, and management practices (Beringer et al., 2011; Kato & Yamagata, 2014; Heck et al.,  
37 2016; Smith et al., 2016; Krause et al., 2017). If woody bioenergy plants replace marginal land, net carbon  
38 uptake increases, enriching soil carbon (Don et al., 2012; Heck et al., 2016; Boysen et al., 2017a,b). On the  
39 other hand, replacing carbon-rich ecosystems with herbaceous bioenergy plants could deplete soil-carbon  
40 stocks and reduce the additional sink capacity of standing forests (Don et al., 2012; Harper et al., 2018).  
41 Furthermore, wood-based BECCS may not be carbon negative in the first decades, initially emitting more  
42 CO<sub>2</sub> than sequestering (Stermann et al., 2018). BECCS has several trade-offs to deal with, including possible  
43 threats to water supply and soil nutrient deficiencies (*medium confidence*) (Smith et al., 2016; Krause et al.,  
44 2017; de Coninck et al., 2018; Heck et al., 2018; Roy et al., 2018) (SRCCL Chapters 2 and 6, Cross-Chapter  
45 Box 5.1). Deployment of BECCS at the scales envisioned by many 1.5–2.0°C mitigation scenarios could  
46 threaten biodiversity and require large land areas, competing with afforestation, reforestation and food  
47 security (Smith et al. 2018; Anderson and Peters 2016). Additional risks and side effects are related to  
48 geologic carbon storage (Fuss et al., 2018) (see also Section 5.6.2.2.4).

49  
50 In conclusion, land-based CDR methods that rely on enhanced net biological uptake and storage of carbon,  
51 have a wide range of biogeochemical and biophysical side effects, which can directly or indirectly strengthen  
52 or weaken the climate mitigation effect of a given method, or affect water quality and quantity, food supply  
53 and biodiversity (Figure 5.36). With the exception of a weakening of ocean carbon sequestration, there is *low*  
54 *confidence* in the Earth system feedbacks of these methods. Most methods are associated with a range of  
55 biogeochemical and biophysical side effects and co-benefits and trade-offs, but these are often highly

1 dependent on local context, management regime, prior land use, and scale (*high confidence*). Highest co-  
2 benefits are obtained with methods that seek to restore natural ecosystems and improve soil carbon  
3 sequestration (Figure 5.36) while highest trade-off possibilities (symmetry with the highest cobenefits) offs  
4 occurs for re/afforestation with monocultures and BECCS, again with strong dependence on scale and  
5 context (*medium confidence*).  
6  
7

#### 8 5.6.2.2.2 *Ocean-based Biological CDR Methods*

9 Both ocean biological and physical processes drive the CO<sub>2</sub> exchange between the ocean and atmosphere.  
10 However, the ocean physical processes that remove CO<sub>2</sub> from the atmosphere, such as large-scale  
11 circulation, cannot be feasibly altered, so ocean CDR methods focus on increasing the productivity of ocean  
12 ecosystems, and subsequent sequestration of carbon (GESAMP, 2019). There has been no change to the  
13 assessment of SROCC (Section 5.5.1): there is *low confidence* that nutrient addition to the open ocean, either  
14 through artificial ocean upwelling or iron fertilisation, could contribute to climate mitigation, due to its  
15 inconclusive effect on carbon sequestration and risks of adverse side effects on marine ecosystems (Figure  
16 5.36; Table 5.9; Supplementary Materials Text 5.SM.3; Supplementary Materials Table 5.SM.4; WGIII,  
17 Section 12.3; Gattuso et al., 2018; Boyd and Vivian, 2019; Feng et al., 2020). In addition, ocean fertilisation  
18 is currently prohibited by the London Protocol (Dixon et al., 2014; GESAMP, 2019).  
19

20 Restoration of vegetated coastal ecosystems (sometimes referred to as ‘blue carbon’ – see Glossary, Annex  
21 VII) refers to the potential for increasing carbon sequestration by plant growth and burial of organic carbon  
22 in the soil of coastal wetlands (including salt marshes and mangroves) and seagrass ecosystems. Wider usage  
23 of the term blue carbon occurs in the literature, for example including seaweeds (macroalgae), shelf sea  
24 sediments and open ocean carbon exchanges. However, such systems are less amenable to management, with  
25 many uncertainties relating to the permanence of their carbon stores (Windham-Myers et al., 2018; Lovelock  
26 and Duarte, 2019; SROCC Section 5.5.1.1).  
27

28 Coastal wetlands and seagrass meadows store significant amounts of carbon and are among the most  
29 productive ecosystems per unit area (Griscom et al., 2017, 2020; Ortega et al., 2019; Serrano et al., 2019).  
30 These rates could be reduced in the future, since these habitats are vulnerable to changing conditions, such as  
31 temperature, salinity, sediment supply, storm severity and continued coastal development (National  
32 Academies of Sciences and Medicine, 2019; Bindoff et al., 2019). These ecosystems are under threat from  
33 anthropogenic conversion and degradation and are being lost at rates between 0.7% and 7% per annum with  
34 consequent CO<sub>2</sub> emissions (e.g. Atwood et al., 2017; Howard et al., 2017; Hamilton and Friess, 2018;  
35 Sasmito et al., 2019). Although sea level rise might lead to greater carbon sequestration in coastal wetlands  
36 (Rogers et al., 2019), there is *high confidence* that the frequency and intensity of marine heatwaves will  
37 increase (Cross-Chapter Box 9.1; Frölicher and Laufkötter, 2018; Laufkötter et al., 2020), which poses a  
38 more immediate threat to the integrity of coastal carbon stocks (Smale et al., 2019). Blue carbon restoration  
39 seeks to increase the rate of carbon sequestration, although restoration may be challenging, because of on-  
40 going use of coastal land for human settlement, conversion to agriculture and aquaculture, shoreline  
41 hardening and port development.  
42

43 Biogeochemical factors affecting reliable quantification of the climatic benefits of coastal vegetation include  
44 the variable production of CH<sub>4</sub> and N<sub>2</sub>O by such ecosystems (Adams et al., 2012; Rosentreter et al., 2018;  
45 Keller, 2019), uncertainties regarding the provenance of the carbon that they accumulate (Macreadie et al.,  
46 2019), and the release of CO<sub>2</sub> by biogenic carbonate formation in seagrass ecosystems (Kennedy et al.,  
47 2018). Whilst coastal habitat restoration potentially provides significant mitigation of national emissions for  
48 some countries (Taillardat et al., 2018; Serrano et al., 2019), the global sequestration potential of blue carbon  
49 approaches is <0.02 PgC yr<sup>-1</sup> (*medium confidence*) (Figure 5.36; SROCC Section 5.5.1.2; Griscom et al.,  
50 2017; Gattuso et al., 2018; National Academies of Sciences and Medicine, 2019).  
51  
52

#### 53 5.6.2.2.3 *Geochemical CDR Methods*

54 Enhanced weathering (EW) is based on naturally occurring weathering processes of silicate and carbonate  
55 rocks, removing CO<sub>2</sub> from the atmosphere. Weathering is accelerated by spreading ground rocks on soils,

1 coasts or oceans. EW increases the alkalinity and pH of natural waters, helps dampen ocean acidification and  
2 increases ocean carbon uptake (Beerling et al., 2018). The dissolution of minerals stimulates biological  
3 productivity of croplands (Hartmann et al., 2013; Beerling et al., 2018), but can also liberate toxic trace  
4 metals (such as Ni, Cr, Cu) into soil or water bodies (Keller et al. 2018; Strefler et al. 2018). EW can also  
5 contribute to freshwater salinisation as a result of increased salt inputs and cation exchange in watersheds,  
6 and so adversely affecting drinking water quality (*low confidence*) (Kaushal et al., 2018). With a *medium*  
7 *confidence*, amendment of soils with minerals will have lower N<sub>2</sub>O emissions (Blanc-Betes et al.; Kantola et  
8 al., 2017) but will not have a marked effect on evapotranspiration or albedo (Fuss et al., 2018; de Oliveira  
9 Garcia et al., 2020). The mining of minerals can cause adverse impacts on biodiversity, however the use of  
10 waste materials such as concrete demolition or steel slags for EW can reduce the need for mining (Renforth,  
11 2019). The spreading of minerals on land has a neutral impact on biodiversity (Smith et al. 2018)

12  
13 Ocean alkalisation, via the deposition of alkaline minerals (e.g. olivine) or their dissociation products (e.g.  
14 quicklime) at the ocean surface, can increase surface total alkalinity and thus increase CO<sub>2</sub> uptake and  
15 storage (see Glossary, Annex VII; Supplementary Material Text 5.SM.3; WGIII Section 12.3; GESAMP  
16 2019; Keller 2019). Ocean alkalisation ameliorates somewhat surface ocean acidification (*high confidence*;  
17 Hauck et al., 2016; Tran et al., 2020), there are also negative side effects on the marine ecosystem, most of  
18 which are poorly understood or quantified (Bach et al., 2019; Figure 5.36; Supplementary Materials Table  
19 5.SM.4). Although ocean alkalisation could potentially sequester large amounts of carbon ( $\geq 1$  PgC yr<sup>-1</sup>;  
20 Figure 5.36; Supplementary Materials Table 5.SM.5) there is no new evidence to revisit the SROCC (Section  
21 5.5.1.2.4) conclusion that there is *low confidence* that ocean alkalisation is a viable climate mitigation  
22 approach.

#### 23 24 25 5.6.2.2.4 Chemical CDR methods

26 Direct air carbon capture with carbon storage (DACCS) is a combination of two techniques, direct capture of  
27 CO<sub>2</sub> from ambient air (DAC) and carbon storage. DAC entails contacting the air, capturing the CO<sub>2</sub> on a  
28 liquid solvent or solid sorbent, and regenerating the solvent or sorbent. Different DAC methods have been  
29 proposed, which differ by the chemical process used to capture the CO<sub>2</sub> and to recover it from the sorbent or  
30 solvent (NASEM, 2019). The captured CO<sub>2</sub> may be either stored geologically as a high-pressure gas or  
31 sequestered by a mineral carbonation process. Storage is potentially permanent in both pressurised gas and  
32 mineral form (Fuss et al., 2018). DACCS has significant requirements of energy and, depending on the type  
33 of technology, water and materials (Smith et al., 2016; NASEM, 2019). Compared to other CDR methods it  
34 has a small land footprint (Smith et al., 2016; NASEM, 2019). Side effects of DACCS include CO<sub>2</sub>-depleted  
35 air leaving the air contactor, which could have adverse effects on crop and ecosystem productivity, and VOC  
36 emissions (NASEM, 2019). Additional risks and side effects are related to the high pressure at which CO<sub>2</sub> is  
37 stored in geologic formations (Fuss et al., 2018). DACCS is assessed in detail in WGIII Section 12.3.

#### 38 39 40 5.6.2.2.5 Methane removal

41 Proposals to remove CH<sub>4</sub> from the atmosphere are emerging (de Richter et al., 2017; Jackson et al., 2019).  
42 CH<sub>4</sub> removal methods seek to capture CH<sub>4</sub> directly from ambient air similarly to DACCS for CO<sub>2</sub> using for  
43 example zeolite trapping, but instead of storing it CH<sub>4</sub> would be chemically oxidized to CO<sub>2</sub> (Jackson et al.,  
44 2019). Methane can be also removed microbially by supporting naturally occurring processes, for example  
45 by enhancing the soil microbial uptake through afforestation (Wu et al., 2018b) or by directing the venting  
46 air from a cow barn into the soil bed of a nearby greenhouse, utilising microbial CH<sub>4</sub> oxidation (Nisbet et al.,  
47 2020). Microbial CH<sub>4</sub> oxidation could also be utilized for removal of CH<sub>4</sub> leaked from point sources by  
48 building biocatalytic polymers which include methane-oxidizing enzymes (Blanchette et al., 2016). Methane  
49 removal is, however, still in its infancy and the available literature is insufficient for an assessment.

50  
51  
52 [START FIGURE 5.36 HERE]

53  
54 **Figure 5.36: Characteristics of carbon dioxide removal (CDR) methods, ordered according to the time scale of**  
55 **carbon storage.** The first column shows biogeophysical (for open-ocean methods) or technical (for all

1 other methods) sequestration potentials (i.e. the sequestration potentials constrained by biological,  
2 geophysical, geochemical limits and thermodynamics and, for technical potentials, availability of  
3 technologies and practices; technical potentials for some methods also consider social or environmental  
4 factors if these represent strong barriers for deployment; see Glossary, Annex VII), classified into low (<  
5  $0.3 \text{ GtCO}_2 \text{ yr}^{-1}$ ), moderate ( $0.3\text{--}3 \text{ GtCO}_2 \text{ yr}^{-1}$ ) and large ( $>3 \text{ GtCO}_2 \text{ yr}^{-1}$ ) (details underlying this  
6 classification are provided in Supplementary Materials Table 5.SM.5). The other columns show Earth  
7 system feedbacks that deployment of a given CDR method would have on carbon sequestration and  
8 climate, along with biogeochemical, biophysical, and other side effects of a given method. Earth system  
9 feedbacks do not include the direct effect of  $\text{CO}_2$  sequestration on atmospheric  $\text{CO}_2$ , only secondary  
10 effects. For Earth system feedbacks, the colours indicate whether the feedbacks strengthen or weaken  
11 carbon sequestration and the climate cooling effect of a given CDR method. For biogeochemical and  
12 biophysical side effects the colours indicate whether the deployment of a CDR method increases or  
13 decreases the magnitude of the effect, whereas for co-benefits and trade-offs the colour indicates whether  
14 deployment of a CDR method results in beneficial (co-benefits) or adverse side-effects (trade-offs) for  
15 water quality and quantity, food production and biodiversity. The details and references underlying the  
16 Earth system feedback and side-effect assessment are provided in Supplementary Materials Table  
17 5.SM.4. Further details on data sources and processing are available in the chapter data table (Table  
18 5.SM.6).

19  
20 **[END FIGURE 5.36 HERE]**

### 21 22 23 **5.6.3 Biogeochemical responses to Solar Radiation Modification (SRM)**

24  
25 This section assesses the possible consequences of solar radiation modification (SRM) on the biosphere and  
26 global biogeochemical cycles. The SRM options (Table 4.6) and the physical climate response to SRM is  
27 assessed in detail in Chapter 4 (Section 4.6.3). Chapter 6 (Section 6.3.6) assesses the potential effective  
28 radiative forcing of aerosol-based SRM options and Chapter 8 (Section 8.6.3) assesses the abrupt water cycle  
29 changes in response to initiation or termination of SRM. Most literature on the biogeochemical responses to  
30 SRM focuses on stratospheric aerosol injection (SAI), and only a few studies have investigated the  
31 biogeochemical responses to marine cloud brightening (MCB) and cirrus cloud thinning (CCT). At the time  
32 of AR5, there were only a few studies on the biogeochemical responses to SRM. The main assessment of  
33 AR5 (Ciais et al., 2013) was that SRM will not interfere with the direct biogeochemical effects of increased  
34  $\text{CO}_2$  such as ocean acidification and  $\text{CO}_2$  fertilisation effect but could affect the carbon cycle through  
35 climate-carbon feedbacks. Overall, AR5 concluded that the level of confidence on the effects of SRM on  
36 carbon and other biogeochemical cycles is *very low* (Ciais et al., 2013). Since AR5, more modelling work  
37 has been conducted to examine various aspects of the global biogeochemical cycle responses to SRM.

#### 38 39 40 **5.6.3.1 Effects of SRM on the Carbon Cycle**

41  
42 Relative to a high-GHG world without SRM, SRM would affect the carbon cycles through changes in  
43 sunlight, climate (e.g. temperature, precipitation, soil moisture, ocean circulation), and atmospheric  
44 chemistry (e.g. ozone) (Cao, 2018) (Section 4.6.3.3). Net SRM effects on the carbon cycle, relative to a  
45 world without SRM, depend on the change of individual factors, and interactions among them.

46  
47 SRM-mediated sunlight changes directly affect the carbon cycle. In particular, SAI would reduce the  
48 sunlight reaching the Earth's surface, but also increase the fraction of sunlight that is diffuse. These changes  
49 in the quantity and quality of the sunlight have opposing effects on the photosynthesis of land plants. On  
50 their own, reductions in photosynthetically active radiation (PAR) will reduce photosynthesis. However,  
51 diffuse light is more effective than direct light in accessing the light-limited leaves within plant canopies,  
52 leading to the so-called 'diffuse-radiation' fertilisation effect (Mercado et al., 2009). The estimated balance  
53 between the negative impacts of reducing PAR and the positive impacts of increasing diffuse fraction differ  
54 between models (Kalidindi et al., 2015; Xia et al., 2016; Yang et al., 2020a) and across different ecosystems.  
55 The change in the absolute amount of direct and diffuse radiation could also depend on the height of the  
56 additional sulphate aerosol layer in the stratosphere and the hygroscopic growth of aerosols (Krishnamohan  
57 et al., 2019, 2020).

1  
2 SRM-mediated cooling also affects the terrestrial carbon cycle. Relative to a high-GHG world without SRM,  
3 the simulated responses of net primary production (NPP) to SRM differ widely between models, such that  
4 even the sign of global mean change is uncertain (Glienke et al., 2015). SRM-induced cooling would  
5 decrease NPP at high latitudes by reducing the length of the growing season (Glienke et al., 2015). At low  
6 latitudes, the NPP response to SRM-induced cooling is sensitive to the effect of nitrogen limitation (Duan et  
7 al., 2020; Glienke et al., 2015; ). SRM-induced cooling tends to increase NPP in models without the nitrogen  
8 cycle because of reduced heat stress. However, in models including the nitrogen cycle, this is counteracted  
9 by reductions in NPP because of reductions in nitrogen mineralisation and nitrogen availability (Glienke et  
10 al., 2015). SRM-induced changes in the hydrological cycle (Section 8.6.3), including changes in  
11 evapotranspiration, precipitation, and soil moisture, also pose strong constraints on the vegetation response  
12 (Dagon and Schrag, 2019). For the same amount of global mean cooling, different SRM options, such as  
13 SAI, MCB, and CCT, would have different effects on GPP and NPP because of different spatial patterns of  
14 temperature available sunlight, and hydrological cycle changes (Section 4.6.3.3) (Duan et al., 2020).  
15 Modelling studies show that SRM-induced cooling would reduce plant and soil respiration (Tjiputra et al.,  
16 2016; Cao and Jiang, 2017; Muri et al., 2018; Yang et al., 2020b). Thus, despite the large uncertainty in  
17 modelled NPP response, existing modelling studies consistently show that relative to a high-CO<sub>2</sub> world  
18 without SRM, SRM would increase the global land carbon sink (*high confidence*).

19  
20 Based on available evidence, SRM with elevated CO<sub>2</sub> would increase global mean NPP and carbon storage  
21 on land, relative to an unperturbed climate mainly because of CO<sub>2</sub> fertilisation of photosynthesis (Glienke et  
22 al., 2015; Tjiputra et al., 2016; Dagon and Schrag, 2019; Duan et al., 2020; Yang et al., 2020b) (*high*  
23 *confidence*). However, the amount of increase is uncertain as it depends on the extent to which CO<sub>2</sub>  
24 fertilisation of land plants is limited by nutrient availability.

25  
26  
27 **[START FIGURE 5.37 HERE]**

28  
29 **Figure 5.37: Cumulative carbon dioxide (CO<sub>2</sub>) uptake by land and ocean carbon sinks in response to**  
30 **stratospheric sulfur dioxide (SO<sub>2</sub>) injection.** Results are shown for a scenario with 50-year  
31 (2020–2069) continuous stratospheric SO<sub>2</sub> injection at a rate of 5 Tg per year applied to a RCP4.5  
32 baseline scenario (GeoMIP experiment G4; Kravitz et al., (2011)), followed by termination in year 2070.  
33 Anomalies are shown relative to RCP4.5 for the multimodel ensemble mean and for each of six ESMs  
34 over the 50-year period of stratospheric SO<sub>2</sub> injection (left), and over 20 years after termination of SO<sub>2</sub>  
35 injection (right). Adapted from Plazzotta et al. (2019). Further details on data sources and processing are  
36 available in the chapter data table (Table 5.SM.6).

37  
38 **[END FIGURE 5.37 HERE]**

39  
40  
41 Relative to a high-CO<sub>2</sub> world without SRM, SRM would also have compensating effects on crop yields.  
42 SRM is expected to have a positive impact on crop yields by diminishing heat stress (Pongratz et al., 2012).  
43 However, reductions in light availability will produce a counteracting reduction in crop yields, especially if  
44 the crop type does not benefit appreciably from diffuse-light fertilisation (Proctor et al., 2018). The balance  
45 between these effects varies markedly across crop types and regions, from projected increases in maize  
46 production in China (Xia et al., 2014) to reductions in groundnut yields in parts of India (Yang et al., 2016).  
47 Because of these diverging results from a limited set of studies, there is overall *low confidence* in the effect  
48 of SRM on crop yields.

49  
50 Consistent with the AR5 assessment, there is *high confidence* that SRM would not mitigate CO<sub>2</sub>-induced  
51 ocean acidification (Ciais et al., 2013). Some studies even suggest an acceleration of deep-ocean  
52 acidification as a result of ocean circulation change (Tjiputra et al., 2016; Lauvset et al., 2017). There are  
53 large differences in the simulated spatial pattern of ocean NPP change in response to SRM, which depends  
54 strongly on the SRM method that is considered (Partanen et al., 2016; Lauvset et al., 2017).

### 5.6.3.2 Consequences of SRM and its termination on atmospheric CO<sub>2</sub> burden

Modelling studies consistently show that relative to a high-CO<sub>2</sub> world without SRM, SRM-induced cooling (Section 4.6.3.3) would reduce plant and soil respiration, and also reduce the negative effects of warming on ocean carbon uptake (Tjiputra et al., 2016; Xia et al., 2016; Cao and Jiang, 2017; Jiang et al., 2018; Muri et al., 2018; Sonntag et al., 2018; Plazzotta et al., 2019; Yang et al., 2020b). A multi-model study (Plazzotta et al., 2019) indicates that relative to a high-CO<sub>2</sub> concentration world without SRM, stratospheric SO<sub>2</sub> injection increases the allowable CO<sub>2</sub> emission by enhancing CO<sub>2</sub> uptake by both land and ocean (Figure 5.37). As a result of enhanced global carbon uptake, SRM would reduce the burden of atmospheric CO<sub>2</sub> (*high confidence*). However, the amount of SRM-induced reduction in atmospheric CO<sub>2</sub> depends on the future emission scenario and modelled oceanic and terrestrial carbon sinks, which differ widely between models (Tjiputra et al., 2016; Xia et al., 2016; Cao and Jiang, 2017; Muri et al., 2018). Models that include the terrestrial nitrogen cycle usually report a much smaller reduction of atmospheric CO<sub>2</sub> in response to SRM than models without the nitrogen cycle mainly because nitrogen limitation leads to a weaker terrestrial carbon sink (Tjiputra et al., 2016; Muri et al., 2018; Yang et al., 2020b). Large scale application of SAI is found to reduce the rate of release of CO<sub>2</sub> and CH<sub>4</sub> from permafrost thaw (Lee et al., 2019; Chen et al., 2020).

A hypothetical, sudden and sustained termination of SRM would cause a rapid increase in global warming that poses great risks to biodiversity (Jones et al., 2013a; McCusker et al., 2014; Trisos et al., 2018) (Section 4.6.3.3). A sudden and sustained termination of SRM would also weaken carbon sinks, accelerating atmospheric CO<sub>2</sub> accumulation and inducing further warming (Figure 5.37) (Matthews and Caldeira, 2007; Tjiputra et al., 2016; Muri et al., 2018; Plazzotta et al., 2019). However, a scenario with gradual phase-out of SRM under emission reduction could reduce the large negative effect of sudden SRM termination (MacMartin et al., 2014; Keith and MacMartin, 2015; Tilmes et al., 2016), though this would be limited by how rapidly emission reductions can be scaled-up (Ekholm and Korhonen, 2016).

### 5.6.3.3 Consequences of SRM on other Biogeochemical Cycles

SAI is found to reduce global average surface ozone concentration (Xia et al., 2017) mainly as a result of aerosol-induced reduction in stratospheric ozone at polar regions, resulting in reduced transport of ozone from the stratosphere (Pitari et al., 2014; Tilmes et al., 2018). The reduction in surface ozone, together with alteration to UV radiation, would have important implications for vegetation response (Xia et al., 2017). A modelling study shows that sea salt aerosol injection for MCB would reduce OH concentration and increase CH<sub>4</sub> lifetime, and hence, have potential implications for surface ozone pollution (Horowitz et al., 2020). It has also been reported that the use of SAI to limit global mean warming from 2°C to 1.5°C would reduce fire weather in many areas (Burton et al., 2018).

### 5.6.3.4 Synthesis of biogeochemical responses to SRM

SRM would alter the global carbon cycle through SRM-induced climate effect such as changes in sunlight, temperature, precipitation, and ocean circulation. Compared to high-CO<sub>2</sub> world without SRM, SRM would enhance the net uptake of CO<sub>2</sub> by the terrestrial biosphere and ocean, thus acting to reduce atmospheric CO<sub>2</sub> (*high confidence*). SRM would also affect surface ozone, UV radiation, and atmospheric chemistry. Due to complex interplays between SRM-induced changes in direct and diffuse sunlight, temperature, the coupling of water-carbon-nitrogen cycles, and atmospheric chemistry, there is large uncertainty in the overall response of the terrestrial biosphere response to SRM. Thus, the level of *confidence* on the effect of SRM on carbon and other biogeochemical cycles is *low*.

## 5.7 Final Remarks

Key research developments to further strengthen the confidence levels in AR7.



### ***Contemporary Greenhouse Gases Trends and Attribution***

- Further constrain the CO<sub>2</sub>, CH<sub>4</sub> and N<sub>2</sub>O fluxes from land use, land use change and forestry (including gross fluxes), and fossil fuels. Improving spatial resolution and representations of land management, such as forestry, grazing and cropping.
- Improve representation of the variability and trends in the transport of carbon through the land to ocean continuum, which has implications for partitioning of the land and ocean CO<sub>2</sub> sinks.
- Improve understanding of the controls over the airborne fraction and sinks rates, their trends, and future dynamics.
- Fill gaps in space and time for ocean CO<sub>2</sub> and ancillary physical and biogeochemical observations at the ocean surface and interior to reduce the biases and uncertainties in the variability and trends for air-sea fluxes and inventory changes, particularly for the Arctic and the Southern Ocean.
- Reduce uncertainties in wetland CH<sub>4</sub> emissions, which is largest source term in the global CH<sub>4</sub> budget and proportionally have the largest uncertainty, in order to better understand future CH<sub>4</sub>-climate feedbacks.
- Reduce uncertainties in atmospheric transport models used to estimate regional sources and sinks of greenhouse gases as independent evidence from that of ground and inventory estimates.

### ***Ocean Acidification and Deoxygenation***

- Improve observations for the interplay between carbonate chemistry and a variety of biogeochemical and physical processes including eutrophication and freshwater inflow in coastal zones to increase the robustness of future assessments of ocean acidification.
- Improve our understanding of changes in water mass ventilation associated with climate change and variability to gain further insights into future trends in ocean acidification and deoxygenation in the ocean interior.

### ***Biogeochemical Feedbacks on Climate Change***

- Improve understanding and representation in Earth system models (ESMs) of changes in land carbon storage and associated carbon-climate feedbacks including better treatment of the CO<sub>2</sub> fertilisation, nutrient-limitations, soil organic matter stabilisation and turnover; land use change; large-scale and fine-scale permafrost carbon; plant growth, mortality, and competition dynamics; plant hydraulics; and disturbance processes.
- Improve observations and process understanding of CH<sub>4</sub> and N<sub>2</sub>O source responses to climate, specifically from wetlands and permafrost thaw.
- Improve observations and process understanding of ocean N<sub>2</sub>O source responses to oxygen loss and climate particularly in the oxygen minimum zones of the tropical oceans and eastern boundary upwelling regions.
- Improve understanding of the sensitivity of ocean carbon-climate feedbacks to physical processes that are not yet resolved by the ocean domain in ESMs.
- Improve understanding of the processes affecting the efficiency, climate sensitivity and emerging feedbacks in the ocean carbon cycle via the biological carbon pump to constrain future global ocean feedbacks.

### ***Remaining Carbon Budget to Climate Stabilisation***

- Improve understanding of both the sign and magnitude of a possible zero emissions commitment (ZEC). The ZEC affects estimates of carbon budgets derived from the TCRE, but not TCRE itself. ZEC is particularly relevant once global CO<sub>2</sub> emissions decline towards net zero.
- Better constrain of the airborne fraction to reduce the spread in TCRE assessment.
- Accounting for timescales of Earth system feedbacks over time for increased accuracy of mitigation needs once global CO<sub>2</sub> emissions reach near-zero levels.

### ***Carbon Dioxide Removal (CDR) and Solar Radiation Modification***

- Run large-scale and long-term experiments and assessments to explore regional feasibility of CDR methods, that they present an actual and verifiable negative regional carbon balance, and that they do not result in adverse unintended consequences.

- 1
- 2
- Improve understanding of the effectiveness of CDR methods to lower atmospheric CO<sub>2</sub> and reduce warming taking into account Earth system feedbacks.

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## Frequently Asked Questions

### FAQ 5.1: Is the natural removal of carbon from the atmosphere weakening?

*For decades, about half of the carbon dioxide (CO<sub>2</sub>) that human activities have emitted to the atmosphere has been taken up by natural carbon sinks in vegetation, soils and oceans. These natural sinks of CO<sub>2</sub> have thus roughly halved the rate at which atmospheric CO<sub>2</sub> concentrations have increased, and therefore slowed down global warming. However, observations show that the processes underlying this uptake are beginning to respond to increasing CO<sub>2</sub> in the atmosphere and climate change in a way that will weaken nature's capacity to take up CO<sub>2</sub> in the future. Understanding of the magnitude of this change is essential for projecting how the climate system will respond to future emissions and emission reduction-efforts.*

Direct observations of CO<sub>2</sub> concentrations in the atmosphere, which began in 1958, show that the atmosphere has only retained roughly half of the CO<sub>2</sub> emitted by human activities due to the combustion of fossil fuels and land-use change such as deforestation (FAQ 5.1, Figure 1). Natural carbon cycle processes on land and in the oceans have taken up the remainder of these emissions. These land and ocean removals or 'sinks' have grown largely in proportion to the increase in CO<sub>2</sub> emissions, taking up 31% (land) and 23% (ocean) of the emissions in 2010–2019, respectively (FAQ 5.1, Figure 1). Therefore, the average proportion of yearly CO<sub>2</sub> emissions staying in the atmosphere has remained roughly stable at 44 % over the last six decades despite continuously increasing CO<sub>2</sub> emissions from human activities.

On land, it is mainly the vegetation that captures CO<sub>2</sub> from the atmosphere through *plant photosynthesis*, which ultimately accumulates both in vegetation and soils. As more CO<sub>2</sub> accumulates in the atmosphere, plant carbon capture increases through the so-called *CO<sub>2</sub> fertilisation effect* in regions where plant growth is not limited by, for instance, nutrient availability. Climate change affects the processes responsible for the uptake and release of CO<sub>2</sub> on land in multiple ways. Land CO<sub>2</sub> uptake is generally increased by longer growing seasons due to global warming in cold regions and by nitrogen deposition in nitrogen-limited regions. Respiration by plants and soil organisms, natural disturbances such as fires, and human activities such as deforestation all release CO<sub>2</sub> back into the atmosphere. The combined effect of climate change on these processes is to weaken the future land sink. In particular, extreme temperatures and droughts as well as permafrost thaw (see FAQ 5.2) tend to reduce the land sink regionally.

In the ocean, several factors control how much CO<sub>2</sub> is captured: the difference in CO<sub>2</sub> partial pressure between the atmosphere and the surface ocean; wind speeds at the ocean surface; the chemical composition of seawater (that is, its *buffering capacity*), which affects how much CO<sub>2</sub> can be taken up; and the use of CO<sub>2</sub> in photosynthesis by seawater microalgae. The CO<sub>2</sub>-enriched surface ocean water is transported to the deep ocean in specific zones around the globe (such as the Northern Atlantic and the Southern Ocean), effectively storing the CO<sub>2</sub> away from the atmosphere for many decades to centuries. The combined effect of warmer surface ocean temperatures on these processes is to weaken the future ocean CO<sub>2</sub> sink.

The ocean carbon sink is better quantified than the land sink thanks to direct ocean and atmospheric carbon observations. The land carbon sink is more challenging to monitor globally, because it varies widely even regionally. There is currently no direct evidence that the natural sinks are slowing down, because observable changes in the fraction of human emissions stored on land or in oceans are small compared to year-to-year and decadal variations of these sinks. Nevertheless, it is becoming more obvious that atmospheric and climate changes are affecting the processes controlling the land and ocean sinks.

Since both the land and ocean sinks respond to the rise in atmospheric CO<sub>2</sub> and to human-induced global warming, the absolute amount of CO<sub>2</sub> taken up by land and ocean will be affected by future CO<sub>2</sub> emissions. This also implies that if countries manage to strongly reduce global CO<sub>2</sub> emissions, or even remove CO<sub>2</sub> from the atmosphere, these sinks will take up less CO<sub>2</sub> because of the reduced human perturbation of the carbon cycle. Under future high-warming scenarios, it is expected that the global ocean and land sinks will stop growing in the second half of the century as climate change increasingly affects them. Thus, both the total amount of CO<sub>2</sub> emitted to the atmosphere and the responses of the natural CO<sub>2</sub> sinks will determine what efforts are required to limit global warming to a certain level (see FAQ 5.4), underscoring how

1 important it is to understand the evolution of these natural CO<sub>2</sub> sinks.  
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4 **[START FAQ 5.1, FIGURE 1 HERE]**  
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6 **FAQ 5.1, Figure 1: Atmospheric CO<sub>2</sub> and natural carbon sinks.** (Top) Global emissions of CO<sub>2</sub> from human  
7 activities and the growth rate of CO<sub>2</sub> in the atmosphere, (middle) the net land and ocean CO<sub>2</sub>  
8 removal (“natural sinks”), as well as (bottom) the fraction of CO<sub>2</sub> emitted by human activities  
9 remaining in atmosphere from 1960 to 2019. Lines are the five years running mean, error-bars  
10 denote the uncertainty of the mean estimate. See Table 5.SM.6 for more information on the data  
11 underlying this figure.  
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13 **[END FAQ 5.1, FIGURE 1 HERE]**  
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**FAQ 5.2: Can thawing permafrost substantially increase global warming?**

*In the Arctic, large amounts of organic carbon are stored in permafrost – ground that remains frozen throughout the year. If significant areas of permafrost thaw as the climate warms, some of that carbon may be released into the atmosphere in the form of carbon dioxide or methane, resulting in additional warming. Projections from models of permafrost ecosystems suggest that future permafrost thaw will lead to some additional warming – enough to be important, but not enough to lead to a ‘runaway warming’ situation, where permafrost thaw leads to a dramatic, self-reinforcing acceleration of global warming.*

The Arctic is the biggest climate-sensitive carbon pool on Earth, storing twice as much carbon in its frozen soils, or *permafrost*, than is currently stored in the atmosphere. As the Arctic region warms faster than anywhere else on earth, there are concerns that this warming could release greenhouse gases to the atmosphere and therefore significantly amplify climate change.

The carbon in the permafrost has built up over thousands of years, as dead plants have been buried and accumulated within layers of frozen soil, where the cold prevents the organic material from decomposing. As the Arctic warms and soils thaw, the organic matter in these soils begins to decompose rapidly and return to the atmosphere as either carbon dioxide or methane, which are both important greenhouse gases. Permafrost can also thaw abruptly in a given place, due to melting ice in the ground reshaping Arctic landscapes, lakes growing and draining, and fires burning away insulating surface soil layers. Thawing of permafrost carbon has already been observed in the Arctic, and climate models project that much of the shallow permafrost (<3 m depth) throughout the Arctic would thaw under moderate to high amounts of global warming (2°C–4°C).

While permafrost processes are complex, they are beginning to be included in models that represent the interactions between the climate and the carbon cycle. The projections from these permafrost carbon models show a wide range in the estimated strength of a carbon–climate vicious circle, from both carbon dioxide and methane, equivalent to 14–175 billion tonnes of carbon dioxide released per 1°C of global warming. By comparison, in 2019, human activities have released about 40 billion tonnes of carbon dioxide into the atmosphere. This has two implications. First, the extra warming caused by permafrost thawing is strong enough that it must be considered when estimating the total amount of remaining emissions permitted to stabilise the climate at a given level of global warming (i.e., the remaining carbon budget, see FAQ 5.4). Second, the models do not identify any one amount of warming at which permafrost thaw becomes a ‘tipping point’ or threshold in the climate system that would lead to a runaway global warming. However, models do project that emissions would continuously increase with warming, and that this trend could last for hundreds of years.

Permafrost can also be found in other cold places (e.g., mountain ranges), but those places contain much less carbon than in the Arctic. For instance, the Tibetan plateau contains about 3% as much carbon as is stored in the Arctic. There is also concern about carbon frozen in shallow ocean sediments. These deposits are known as *methane hydrates* or *clathrates*, which are methane molecules locked within a cage of ice molecules. They formed as frozen soils that were flooded when sea levels rose after the last ice age. If these hydrates thaw, they may release methane that can bubble up to the surface. The total amount of carbon in permafrost-associated methane hydrates is much less than the carbon in permafrost soils. Global warming takes millennia to penetrate into the sediments beneath the ocean, which is why these hydrates are still responding to the last deglaciation. As a result, only a small fraction of the existing hydrates could be destabilised during the coming century. Even when methane is released from hydrates, most of it is expected to be consumed and oxidised into carbon dioxide in the ocean before reaching the atmosphere. The most complete modelling of these processes to date suggests a release to the atmosphere at a rate of less than 2% of current human-induced methane emissions.

Overall, thawing permafrost in the Arctic appears to be an important additional source of heat-trapping gases to the atmosphere, more so than undersea hydrates. Climate and carbon cycle models are beginning to consider permafrost processes. While these models disagree on the exact amount of the heat-trapping gases that will be released into the atmosphere, they agree (i) that the amount of such gases released from

1 permafrost will increase with the amount of global warming, and (ii) that the warming effect of thawing  
2 permafrost is significant enough to be considered in estimates of the remaining carbon budgets for limiting  
3 future warming.  
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6 **[START FAQ 5.2, FIGURE 1 HERE]**  
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8 **FAQ 5.2, Figure 1: The Arctic permafrost is a big pool of carbon that is sensitive to climate change.** (left)  
9 Quantity of carbon stored in the permafrost, to 3 m depth (NCSCDV2 dataset) and (right) area of  
10 permafrost vulnerable to abrupt thaw (Circumpolar Thermokarst Landscapes dataset).  
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12 **[END FAQ 5.2, FIGURE 1 HERE]**  
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**FAQ 5.3: Could climate change be reversed by removing carbon dioxide from the atmosphere?**

*Deliberate removal of carbon dioxide (CO<sub>2</sub>) from the atmosphere could reverse (i.e. change the direction of) some aspects of climate change. However, this will only happen if it results in a net reduction in the total amount of CO<sub>2</sub> in the atmosphere, that is, if deliberate removals are larger than emissions. Some climate change trends, such as the increase in global surface temperature, would start to reverse within a few years. Other aspects of climate change would take decades (e.g., permafrost thawing) or centuries (e.g., acidification of the deep ocean) to reverse, and some, such as sea level rise, would take centuries to millennia to change direction.*

The term *negative carbon dioxide (CO<sub>2</sub>) emissions* refers to the removal of CO<sub>2</sub> from the atmosphere by deliberate human activities, in addition to the removals that occur naturally, and is often used as synonymous with *carbon dioxide removal*. Negative CO<sub>2</sub> emissions can compensate for the release of CO<sub>2</sub> into the atmosphere by human activities. They could be achieved by strengthening natural CO<sub>2</sub> sequestration processes on land (e.g., by planting trees or through agricultural practices that increase the carbon content of soils) and/or in the ocean (e.g., by restoration of coastal ecosystems) or by removing CO<sub>2</sub> directly from the atmosphere. If CO<sub>2</sub> removals are greater than human-caused CO<sub>2</sub> emissions globally, emissions are said to be *net negative*. It should be noted that CO<sub>2</sub> removal technologies are not yet ready or unable to achieve the scale of removal that would be required to compensate for current levels of emissions, and most have undesired side effects.

In the absence of deliberate CO<sub>2</sub> removal, the CO<sub>2</sub> concentration in the atmosphere (a measure of the amount of CO<sub>2</sub> in the atmosphere) results from a balance between human-caused CO<sub>2</sub> release and the removal of CO<sub>2</sub> by natural processes on land and in the ocean (natural ‘carbon sinks’) (see FAQ 5.1). If CO<sub>2</sub> release exceeds removal by carbon sinks, the CO<sub>2</sub> concentration in the atmosphere would increase; if CO<sub>2</sub> release equals removal, the atmospheric CO<sub>2</sub> concentration would stabilise; and if CO<sub>2</sub> removal exceeds release, the CO<sub>2</sub> concentration would decline. This applies in the same way to *net* CO<sub>2</sub> emissions that is, the sum of human-caused releases and deliberate removals.

If the CO<sub>2</sub> concentration in the atmosphere starts to go down, the Earth’s climate would respond to this change (FAQ 5.3, Figure 1). Some parts of the climate system take time to react to a change in CO<sub>2</sub> concentration, so a decline in atmospheric CO<sub>2</sub> as a result of net negative emissions would not lead to immediate reversal of all climate change trends. Recent studies have shown that global surface temperature starts to decline within a few years following a decline in atmospheric CO<sub>2</sub>, although the decline would not be detectable for decades due to natural climate variability (see FAQ 4.2). Other consequences of human-induced climate change such as reduction in permafrost area would take decades, and yet others such as warming, acidification and oxygen loss of the deep ocean would take centuries to reverse following a decline in the atmospheric CO<sub>2</sub> concentration. Sea level would continue to rise for many centuries to millennia, even if large deliberate CO<sub>2</sub> removals were successfully implemented.

A class of future scenarios that is receiving increasing attention, particularly in the context of ambitious climate goals such as the global warming limits of 1.5°C or 2°C included in the Paris Agreement, are so-called ‘overshoot’ scenarios. In these scenarios, a slow rate of reductions in emissions in the near term is compensated by net negative CO<sub>2</sub> emissions in the later part of this century, which results in a temporary breach or ‘overshoot’ of a given warming level. Due to the delayed reaction of several climate system components, it follows that the temporary overshoot would result in additional climate changes compared to a scenario that reaches the goal without overshoot. These changes would take decades to many centuries to reverse, with the reversal taking longer for scenarios with larger overshoot.

Removing more CO<sub>2</sub> from the atmosphere than is emitted into it would indeed begin to reverse some aspects of climate change, but some changes would still continue in their current direction for decades to millennia. Approaches capable of large-scale removal of CO<sub>2</sub> are still in the state of research and development or unproven at the scales of deployment necessary to achieve a net reduction in atmospheric CO<sub>2</sub> levels. CO<sub>2</sub> removal approaches, particularly those deployed on land, can have undesired side-effects on water, food production and biodiversity.

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[START FAQ 5.3, FIGURE 1 HERE]

**FAQ 5.3, Figure 1: Changes in aspects of climate change in response to a peak and decline in the atmospheric CO<sub>2</sub> concentration** (top panel). The vertical grey dashed line indicates the time of peak CO<sub>2</sub> concentration in all panels. It is shown that the reversal of global surface warming lags the decrease in the atmospheric CO<sub>2</sub> concentration by a few years, the reversal of permafrost area decline lags the decrease in atmospheric CO<sub>2</sub> by decades, and ocean thermal expansion continues for several centuries. Note that the quantitative information in the figure (i.e., numbers on vertical axes) is not to be emphasized as it results from simulations with just one model and will be different for other models. The qualitative behaviour, however, can be expected to be largely model independent.

[END FAQ 5.3, FIGURE 1 HERE]

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**FAQ 5.4: What are carbon budgets?**

There are several types of carbon budgets. Most often, the term refers to the total net amount of carbon dioxide (CO<sub>2</sub>) that can still be emitted by human activities while limiting global warming to a specified level (e.g., 1.5°C or 2°C above pre-industrial levels). This is referred to as the ‘remaining carbon budget’. Several choices and value judgments have to be made before it can be unambiguously estimated. When the remaining carbon budget is combined with all past CO<sub>2</sub> emissions to date, a ‘total carbon budget’ compatible with a specific global warming limit can also be defined. A third type of carbon budget is the ‘historical carbon budget’, which is a scientific way to describe all past and present sources and sinks of CO<sub>2</sub>.

The term *remaining carbon budget* is used to describe the total net amount of CO<sub>2</sub> that human activities can still release into the atmosphere while keeping global warming to a specified level, like 1.5°C or 2°C relative to pre-industrial temperatures. Emissions of CO<sub>2</sub> from human activities are the main cause of global warming. A remaining carbon budget can be defined because of the specific way CO<sub>2</sub> behaves in the Earth system. That is, global warming is roughly linearly proportional to the total net amount of CO<sub>2</sub> emissions that are released into the atmosphere by human activities – also referred to as cumulative anthropogenic CO<sub>2</sub> emissions. Other greenhouse gases behave differently and have to be accounted for separately.

The concept of a remaining carbon budget implies that to stabilize global warming at any particular level, global emissions of CO<sub>2</sub> need to be reduced to net zero levels at some point. Net zero CO<sub>2</sub> emissions describes a situation in which all the anthropogenic emissions of CO<sub>2</sub> are counterbalanced by deliberate anthropogenic removals so that on average no CO<sub>2</sub> is added or removed from the atmosphere by human activities. Atmospheric CO<sub>2</sub> concentrations in such a situation would gradually decline to a long-term stable level as excess CO<sub>2</sub> in the atmosphere is taken up ocean and land sinks (see FAQ 5.1). The concept of a remaining carbon budget also means that if CO<sub>2</sub> emissions reductions are delayed, deeper and faster reductions are needed later to stay within the same budget. If the remaining carbon budget is exceeded, this will result in either higher global warming or a need to actively remove CO<sub>2</sub> from the atmosphere to reduce global temperatures back down to the desired level (see FAQ 5.3).

Estimating the size of remaining carbon budgets depends on a set of choices. These choices include (1) the global warming level that is chosen as a limit (for example, 1.5°C or 2°C relative to pre-industrial levels), (2) the probability with which we want to ensure that warming is held below that limit (for example, a one-in-two, two-in-three, or higher chance), and (3) how successful we are in limiting emissions of other greenhouse gases that affect the climate, such as methane or nitrous oxide. These choices can be informed by science but ultimately represent subjective choices. Once they have been made, we can combine knowledge about how much our planet has warmed already, about the amount of warming per cumulative tonne of CO<sub>2</sub>, and about the amount of warming that is still expected once global net CO<sub>2</sub> emissions are brought down to zero to estimate the remaining carbon budget for a given temperature goal. For example, to limit global warming to 1.5°C above pre-industrial levels with either a one-in-two (50%) or two-in-three (67%) chance, the remaining carbon budgets amount to 500 and 400 billion tonnes of CO<sub>2</sub>, respectively, from 1 January 2020 onward (FAQ 5.4, Figure 1). Currently, human activities are emitting around 40 billion tonnes of CO<sub>2</sub> into the atmosphere in a single year.

The remaining carbon budget depends on how much the world has already warmed to date. This past warming is caused by historical emissions, which are estimated by looking at the *historical carbon budget* – a scientific way to describe all past and present sources and sinks of CO<sub>2</sub>. It describes how the CO<sub>2</sub> emissions from human activities have redistributed across the various CO<sub>2</sub> reservoirs of the Earth system. These reservoirs are the ocean, the land vegetation, and the atmosphere (into which CO<sub>2</sub> was emitted). The share of CO<sub>2</sub> that is not taken up by the ocean or the land, and that thus increases the concentration of CO<sub>2</sub> in the atmosphere, causes global warming. The historical carbon budget tells us that of the about 2560 billion tonnes of CO<sub>2</sub> that were released into the atmosphere by human activities between the years 1750 and 2019, about a quarter were absorbed by the ocean (causing ocean acidification) and about a third by the land vegetation. About 45% of these emissions remain in the atmosphere (see FAQ 5.1). Adding these historical CO<sub>2</sub> emissions to estimates of remaining carbon budgets allows one to estimate the *total carbon budget*

1 consistent with a specific global warming level.

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3 In summary, determining a remaining carbon budget – that is, how much CO<sub>2</sub> can be released into the  
4 atmosphere while stabilizing global temperature below a chosen level – is well understood but relies on a set  
5 of choices. What is clear, however, is that for limiting warming below 1.5°C or 2°C the remaining carbon  
6 budget from 2020 onwards is much smaller than the total CO<sub>2</sub> emissions released to date.

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8 **[START FAQ 5.4, FIGURE 1]**

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10 **FAQ 5.4, Figure 1: Various types of carbon budgets.** Historical cumulative CO<sub>2</sub> emissions determine to a large  
11 degree how much the world has warmed to date, while the remaining carbon budget indicates how much  
12 CO<sub>2</sub> could still be emitted while keeping warming below specific temperature thresholds. Several factors  
13 limit the precision with which the remaining carbon budget can be estimated, and estimates therefore need to  
14 specify the probability with which they aim at limiting warming to the intended target level (e.g., limiting  
15 warming to 1.5°C with a 67% probability).

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18 **[END FAQ 5.4, FIGURE 1]**

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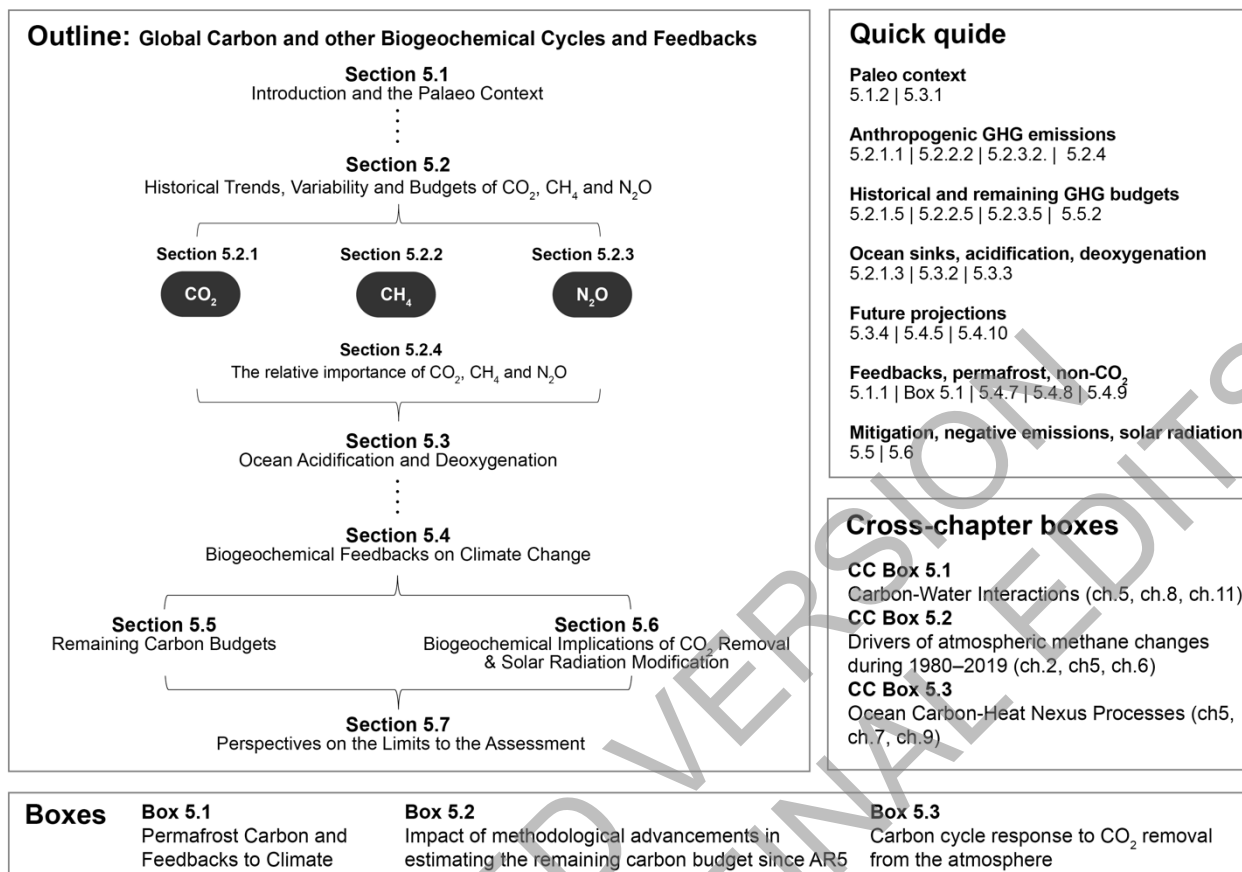
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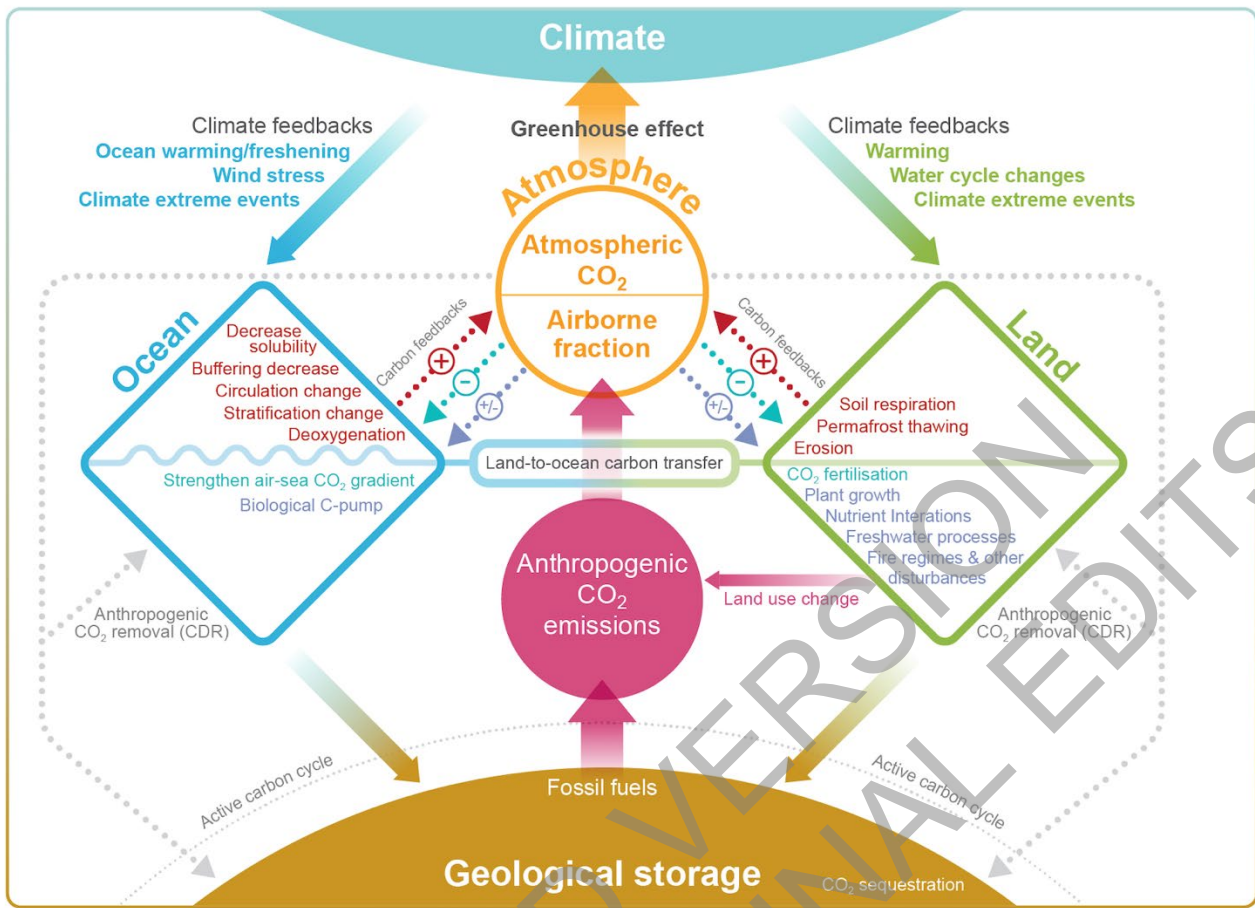
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**Figure 5.1:**

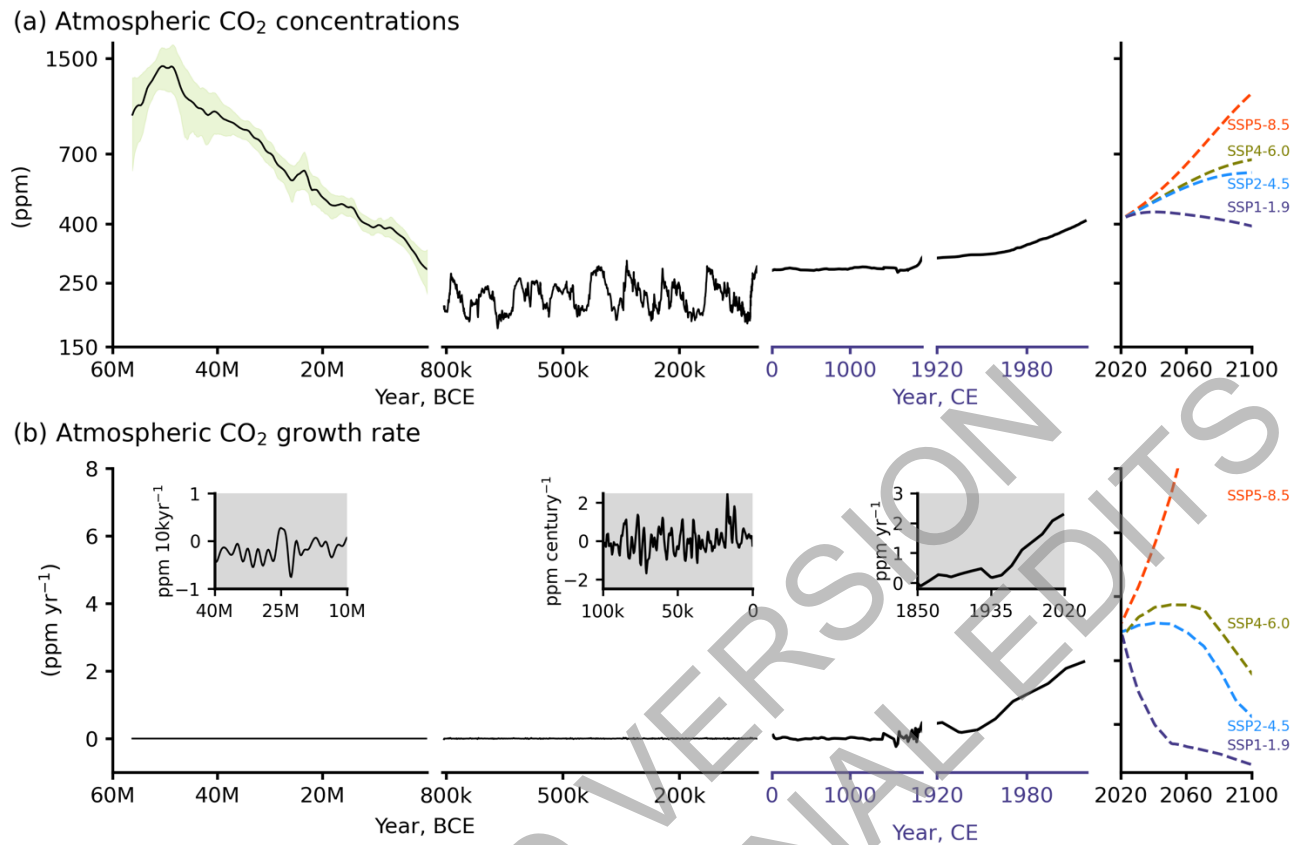
**Visual abstract for Chapter 5**





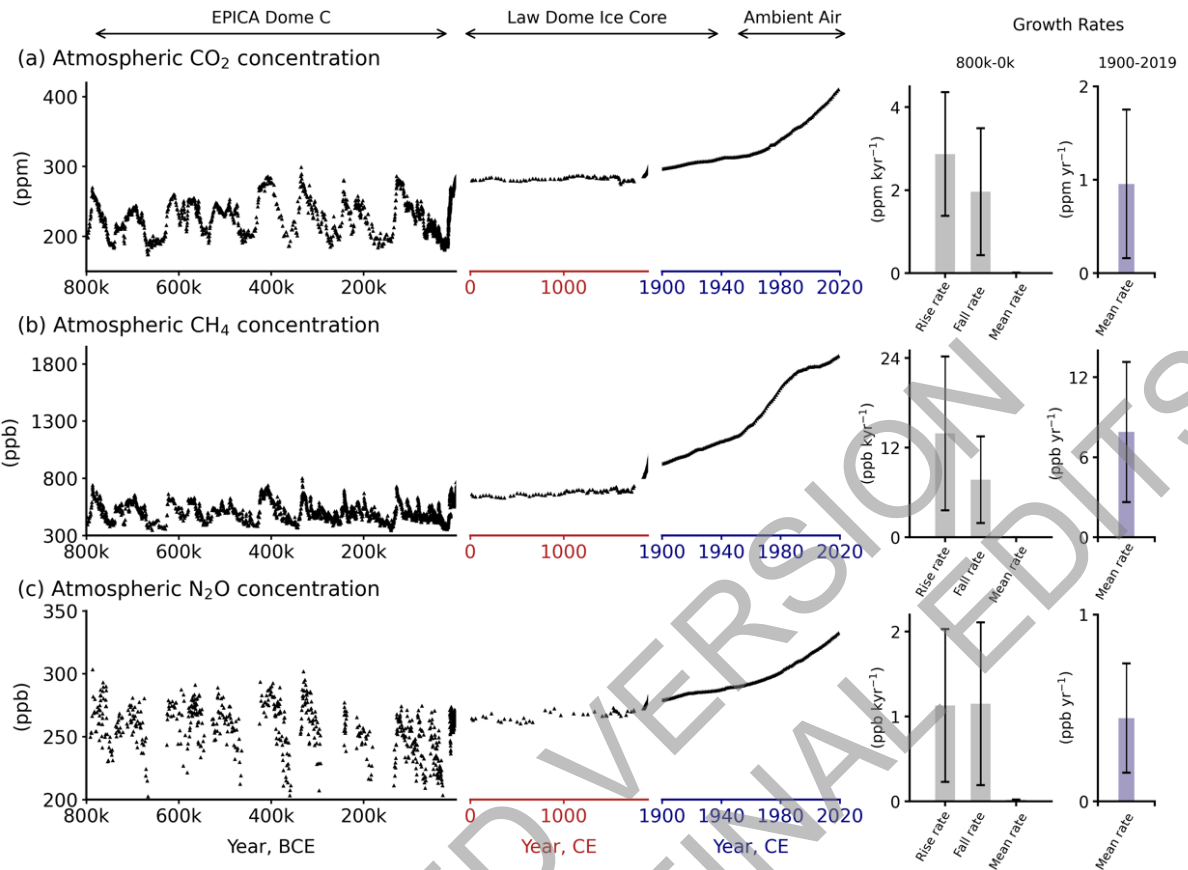
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**Figure 5.2: Key compartments, processes and pathways that govern historical and future CO<sub>2</sub> concentrations and carbon–climate feedbacks through the coupled earth system.** The anthropogenic CO<sub>2</sub> emissions, including land use change, are partitioned via negative feedbacks (turquoise dotted arrows) between the ocean (23%), the land (31%) and the airborne fraction (46%) of anthropogenic CO<sub>2</sub> that sets the changing CO<sub>2</sub> concentration in the atmosphere (2010–2019, Table 5.1). This regulates most of the radiative forcing that creates the heat imbalance that drives the climate feedbacks to the ocean (blue) and land (green). Positive feedbacks (red arrows) result from processes in the ocean and on land (red text). Positive feedbacks are influenced by both carbon-concentration and carbon-climate feedbacks simultaneously. Additional biosphere processes have been included but these have an as yet uncertain feedback impact (blue-dotted arrows). CO<sub>2</sub> removal from the atmosphere into the ocean, land and geological reservoirs, necessary for negative emissions, has been included (grey arrows). Although this schematic is built around CO<sub>2</sub>, the dominant GHG, some of the same processes also influence the fluxes of CH<sub>4</sub> and N<sub>2</sub>O and the strength of the positive feedbacks from the terrestrial and ocean systems.

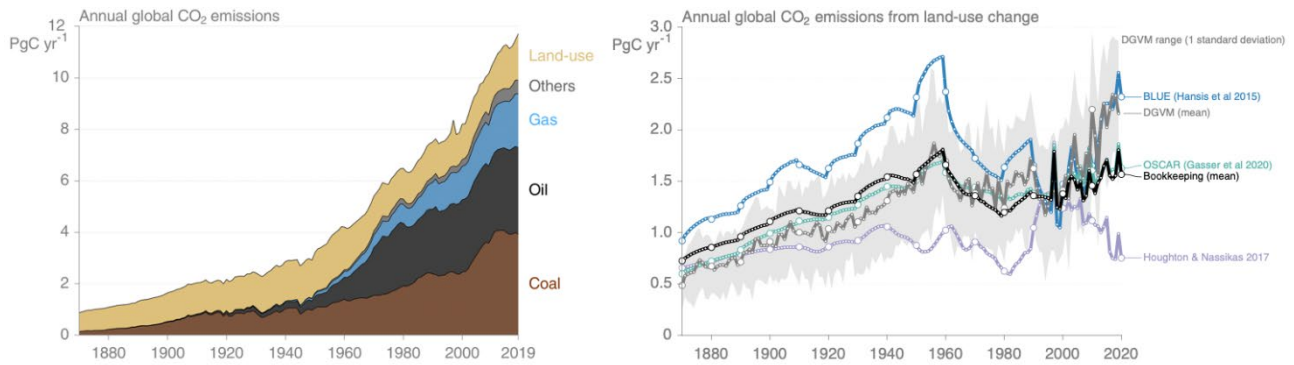


**Figure 5.3: Atmospheric CO<sub>2</sub> concentrations and growth rates for the past 60 million years and projections to 2100.** (a) CO<sub>2</sub> concentrations. Concentrations data for the period 60 Myr to the time prior to 800 Kyr (left column) are shown as the LOESS Fit and 68% range (data from Chapter 2) (Foster et al., 2017). Concentrations from 1750 and projections through 2100 are taken from Shared Socioeconomic Pathways of IPCC AR6 (Meinshausen et al., 2017). (b) Growth rates are shown as the time derivative of the concentration time series. Inserts in (b) show growth rates at the scale of the sampling resolution. Further details on data sources and processing are available in the chapter data table (Table 5.SM.6).

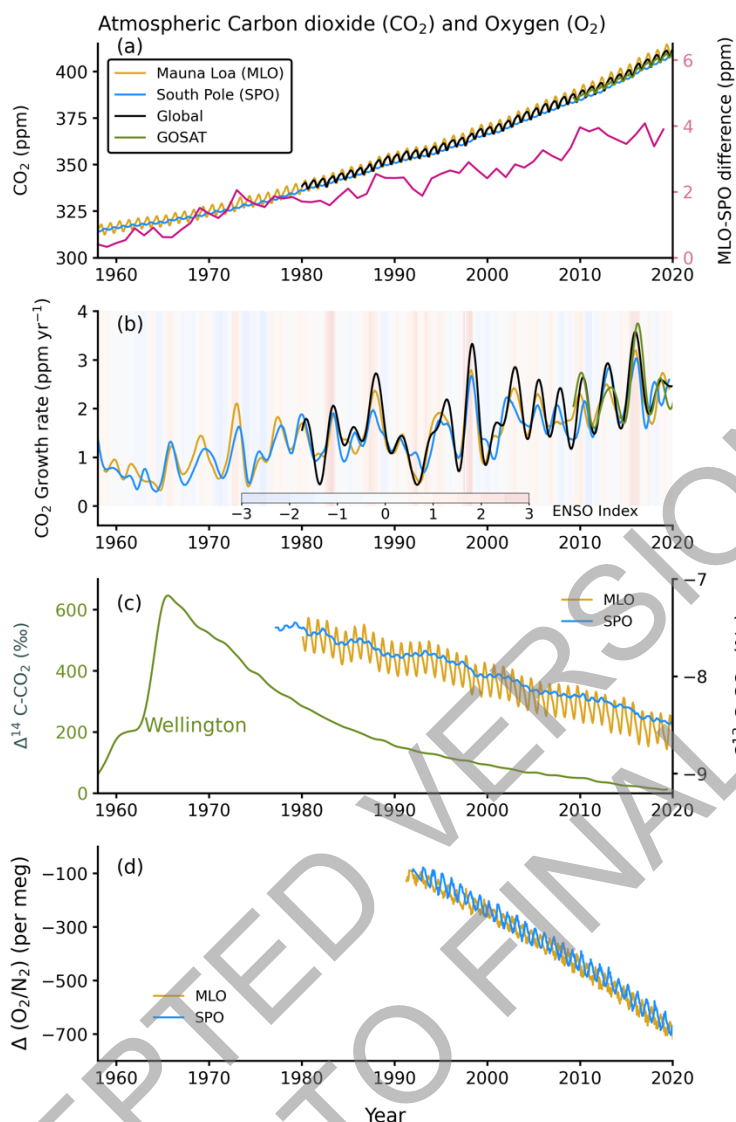
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**Figure 5.4: Atmospheric concentrations of CO<sub>2</sub>, CH<sub>4</sub> and N<sub>2</sub>O in air bubbles and clathrate crystals in ice cores (800,000 BCE to 1990 CE).** Note the variable x-axis range and tick mark intervals for the 3 columns. Ice core data is over-plotted by atmospheric observations from 1958 to present for CO<sub>2</sub>, from 1984 for CH<sub>4</sub> and from 1994 for N<sub>2</sub>O. The time-integrated, millennial-scale linear growth rates for different time periods (800,000–0 BCE, 0–1900 CE and 1900–2017 CE) are given in each panel. For the BCE period, mean rise and fall rates are calculated for the individual slopes between the peaks (interglacials) and troughs (ice ages), which are given in the panels in left column. The data for BCE period are used from the Vostok, EPICA, Dome C and WAIS ice cores (Petit et al., 1999; Monnin, 2001; Pépin et al., 2001; Raynaud et al., 2005; Siegenthaler et al., 2005; Loulergue et al., 2008; Lüthi et al., 2008; Schilt et al., 2010a). The data after 0-yr CE are taken mainly from Law Dome ice core analysis (MacFarling Meure et al., 2006). The surface observations for all species are taken from NOAA cooperative research network (Dlugokencky and Tans, 2019), where ALT, MLO and SPO stand for Alert (Canada), Mauna Loa Observatory, and South Pole Observatory, respectively. BCE = before current era, CE = current era. Further details on data sources and processing are available in the chapter data table (Table 5.SM.6).

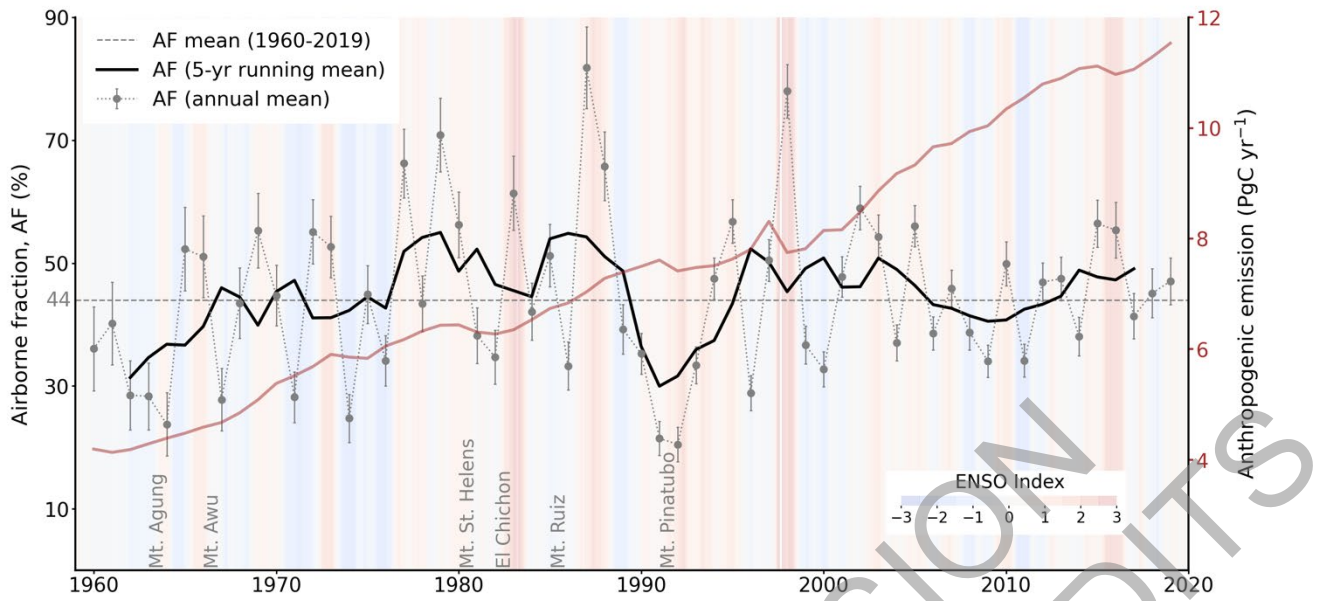


**Figure 5.5: Global anthropogenic CO<sub>2</sub> emissions.** (a) Historical trends of anthropogenic CO<sub>2</sub> emission (fossil fuels and net land use change, including land management, called LULUCF flux in the main text) for the period 1870 to 2019, with ‘others’ representing flaring, emissions from carbonates during cement manufacture. Data sources: (Boden et al., 2017; IEA, 2017; Andrew, 2018; BP, 2018; Le Quéré et al., 2018; Friedlingstein et al., 2020). (b) The net land use change CO<sub>2</sub> flux (PgC yr<sup>-1</sup>) as estimated by three bookkeeping models and 16 Dynamic Global Vegetation Models (DGVMs) for the global annual carbon budget 2019 (Friedlingstein et al., 2020). The three bookkeeping models are from Hansis et al., 2015; Houghton and Nassikas, 2017; (Gasser et al., 2020) and are all updated to 2019; their average is used to determine the net land use change flux in the annual global carbon budget (black line). The DGVM estimates are the result of differencing a simulation with and without land use changes run under observed historical climate and CO<sub>2</sub>, following the Trendy v9 protocol; they are used to provide an uncertainty range to the bookkeeping estimates (Friedlingstein et al., 2020). All estimates are unsmoothed annual data. Estimates differ in process comprehensiveness of the models and in definition of flux components included in the net land use change flux. Further details on data sources and processing are available in the chapter data table (Table 5.SM.6).



**Figure 5.6: Time series of CO<sub>2</sub> concentrations and related measurements in ambient air.** (a) concentration time series and MLO-SPO difference, (b) growth rates, (c) <sup>14</sup>C and <sup>13</sup>C isotopes, and (d) O<sub>2</sub>/N<sub>2</sub> ratio. The data for Mauna Loa Observatory (MLO) and South Pole Observatory (SPO) are taken from the Scripps Institution of Oceanography (SIO)/University of California, San Diego (Keeling et al., 2001). The global mean CO<sub>2</sub> are taken from NOAA cooperative network (as in Chapter 2), and GOSAT monthly-mean XCO<sub>2</sub> time series are taken from National Institute for Environmental Studies (Yoshida et al., 2013). CO<sub>2</sub> growth rates are calculated as the time derivative of deseasonalised time series (Nakazawa et al., 1997). The Δ(O<sub>2</sub>/N<sub>2</sub>) are expressed in per meg units (= (FF/M)×10<sup>6</sup>, where FF = moles of O<sub>2</sub> consumed by fossil-fuel burning, M = 3.706×10<sup>19</sup>, total number of O<sub>2</sub> molecules in the atmosphere (Keeling and Manning, 2014). The <sup>14</sup>CO<sub>2</sub> time series at Barring Head, Wellington, New Zealand (BHD) is taken from GNS Science and NIWA (Turnbull et al., 2017). The multivariate ENSO index (MEI) is shown as the shaded background in panel (b; warmer shade indicates El Niño). Further details on data sources and processing are available in the chapter data table (Table 5.SM.6).

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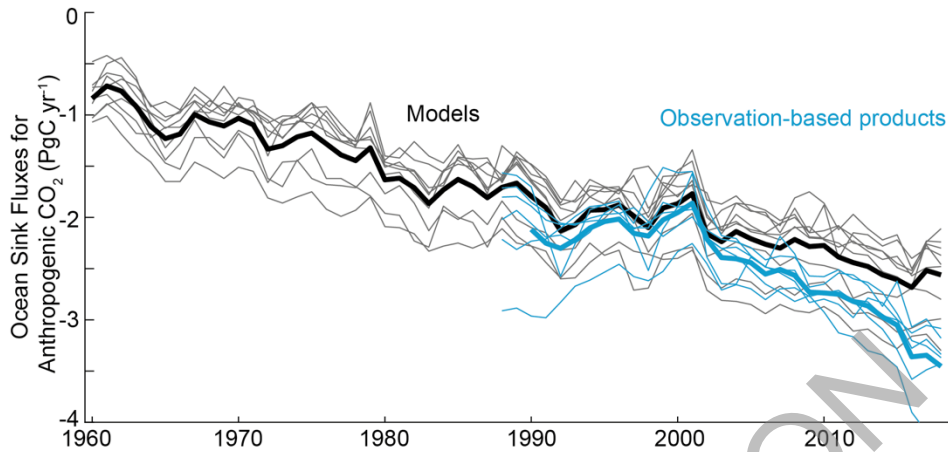


**Figure 5.7: Airborne fraction and anthropogenic (fossil fuel and land use change) CO<sub>2</sub> emissions.** Data as in Section 5.2.1.1. The multivariate ENSO index (shaded) and the major volcanic eruptions are marked along the x-axis. Further details on data sources and processing are available in the chapter data table (Table 5.SM.6).

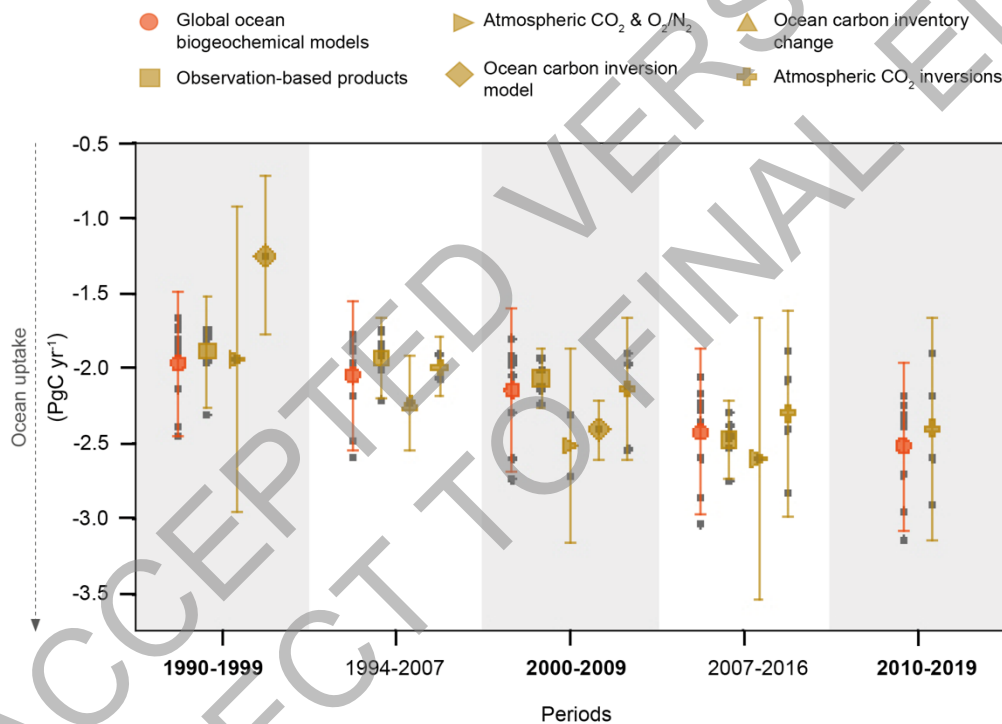
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(a) Comparative trends in the ocean sink for anthropogenic CO<sub>2</sub>: models (1960–2018) and observation-based products (1987–2018)



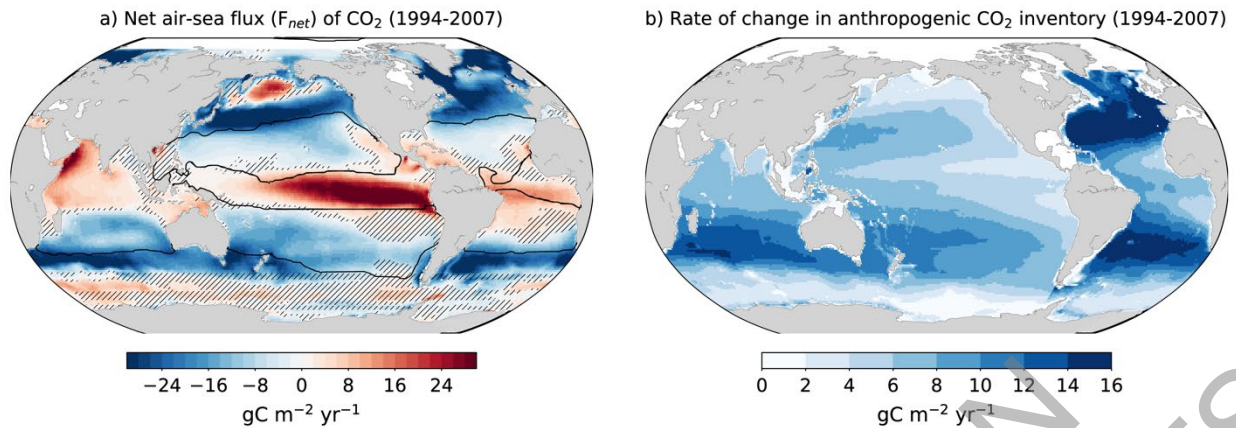
(b) Multi-method Decadal means of the Ocean Sink Fluxes for Anthropogenic CO<sub>2</sub>



**Figure 5.8: Multi-decadal trends for the ocean sink of CO<sub>2</sub>.** (a): The multi-decadal (1960–2019) trends in the annual ocean sink ( $S_{\text{ocean}}$ ) reconstructed from 9 Global Ocean Biogeochemical Models (GOBM) forced with observed atmospheric re-analysis products (Hauck et al., 2020), 6 observationally-based gap-filling products that reconstructed spatial and temporal variability in the ocean CO<sub>2</sub> flux from sparse observations of surface ocean  $p\text{CO}_2$  (Supplementary Materials 5.SM.2). The trends in  $S_{\text{ocean}}$  were calculated from the mean annual global GOBM outputs and the observational products were used to provide confidence in the GOBM assessments ( $r^2=0.85$ ). Thick lines represent the multi-model mean. Observationally-based products have been corrected for pre-industrial river carbon fluxes (0.62 PgC yr<sup>-1</sup>) based on the average of estimates from Jacobson et al., (2007) and Resplandy et al. (2018). (b): Mean decadal constraints and their confidence intervals for global ocean sink ( $S_{\text{ocean}}$ ) of anthropogenic CO<sub>2</sub> using multiple independent or quasi-independent lines of evidence or methods for the period 1990–2019 (See Supplementary Materials Tables 5.SM.1 and 5.SM.2 for magnitudes, uncertainties and published sources). Further details on data sources and processing are available in the chapter data table (Table 5.SM.6).

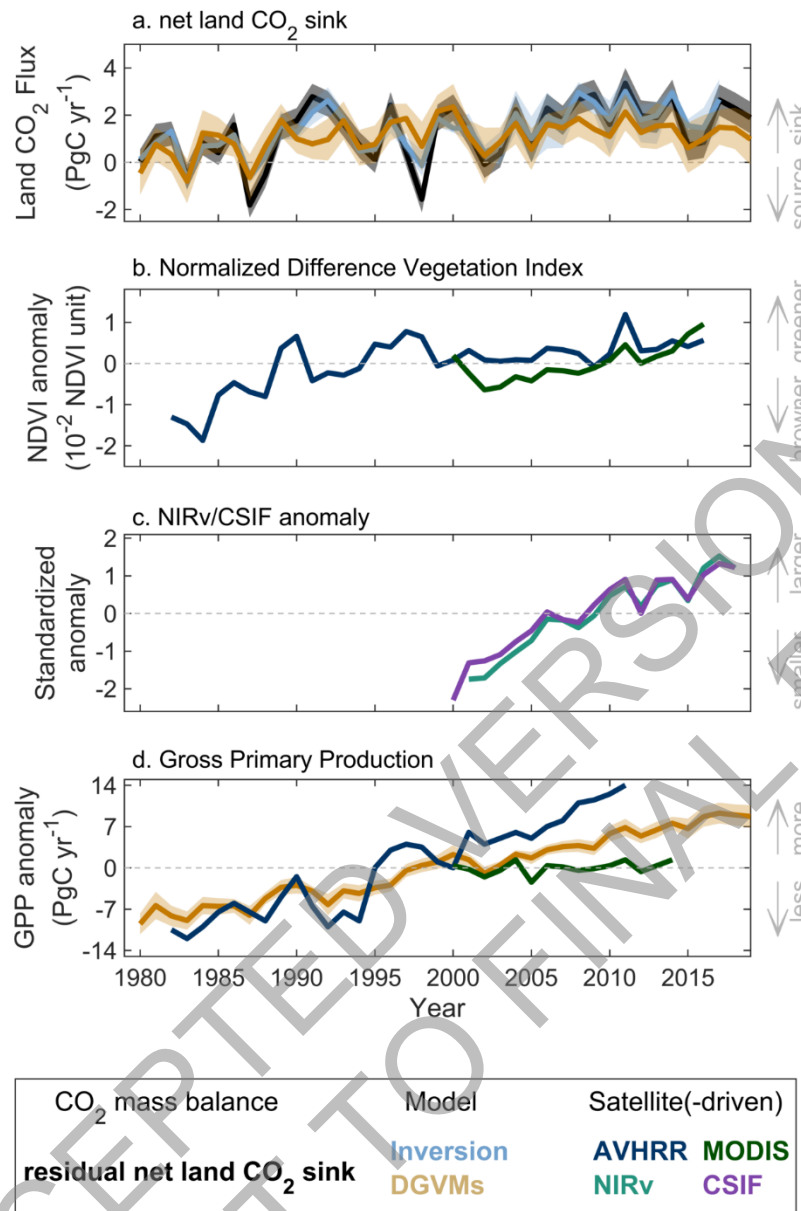
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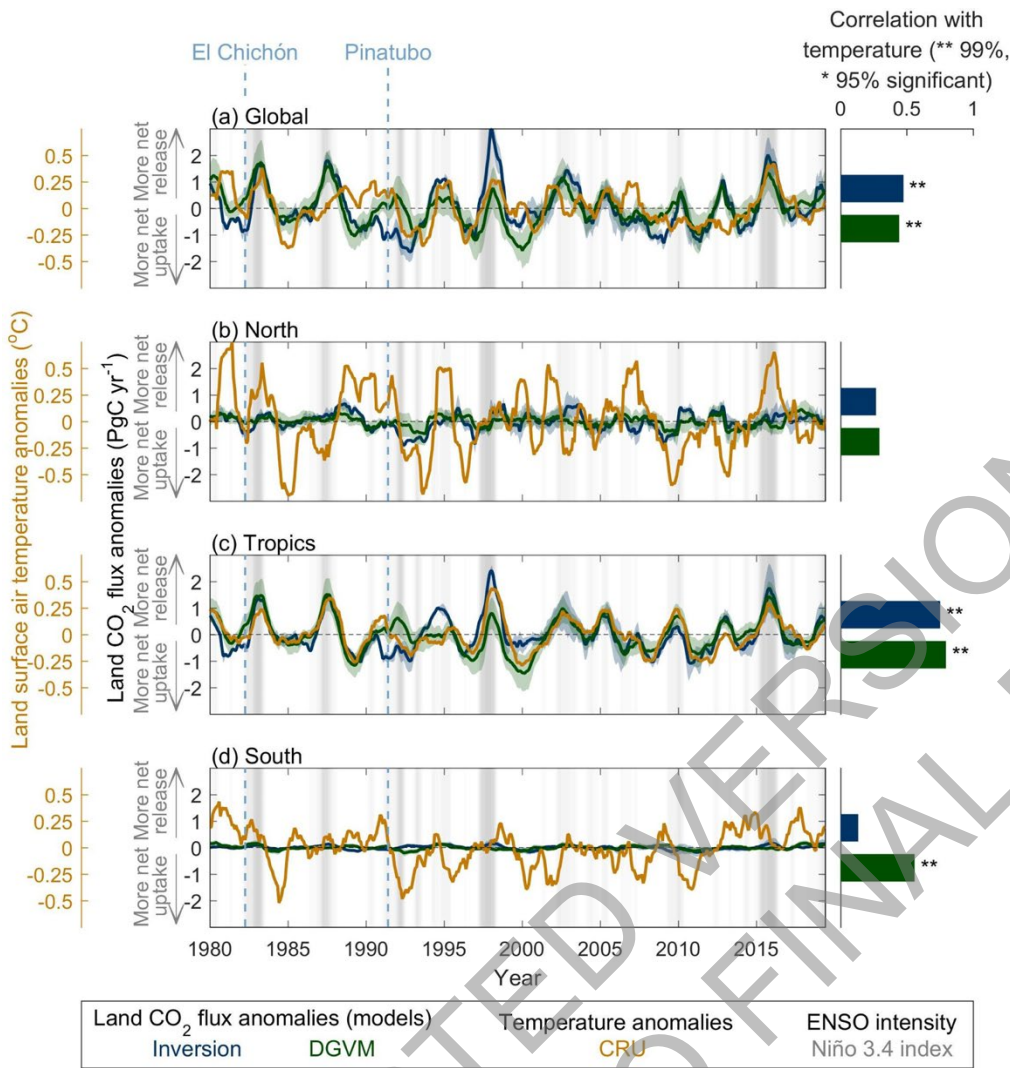
**Figure 5.9: Comparative regional characteristics of the mean decadal (1994–2007) sea-air CO<sub>2</sub> flux ( $F_{net}$ ) and ocean storage of anthropogenic CO<sub>2</sub>.** (a) regional sink–source characteristics for contemporary ocean air-sea CO<sub>2</sub> fluxes ( $F_{net}$ ) derived from the ensemble of 6 observation-based products using SOCATv6 observational data set (Bakker et al., 2020; Landschützer et al., 2014; Rödenbeck et al., 2014; Zeng et al., 2014; Denvil-Sommer et al., 2019; Gregor et al., 2019; Iida et al., 2020) Warm colours depict outgassing fluxes and black contours characterise the super-biomes defined from Fay and McKinley (2014) and adjusted by Gregor et al., (2019) also used to calculate the variability in regional flux anomalies (Supplementary Materials Figure 5.SM.1); and (b) the regional characteristics of the storage fluxes of CO<sub>2</sub> in the ocean interior for the same period (Gruber et al., 2019a). The dots reflect ocean areas where the 1-sigma standard deviation of  $F_{net}$  from the 6 observation product is larger than the magnitude of the mean. This reflects source-sink transition areas where the mean  $F_{net}$  is small and more strongly influenced by spatial and temporal variability across the products. Further details on data sources and processing are available in the chapter data table (Table 5.SM.6).





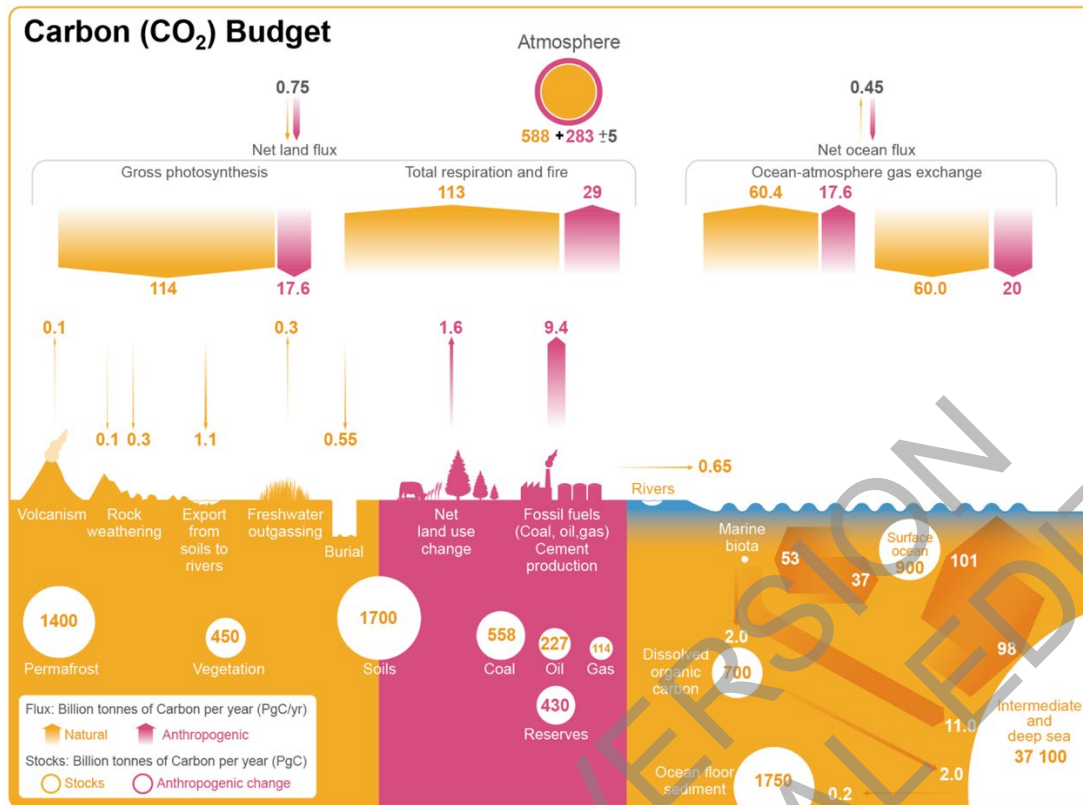
**Figure 5.10: Trends of the net land CO<sub>2</sub> sink and related vegetation observations during 1980–2019.** (a) Net land CO<sub>2</sub> sink. The residual net land CO<sub>2</sub> sink is estimated from the global CO<sub>2</sub> mass balance (fossil fuel emission minus atmospheric CO<sub>2</sub> growth rate and ocean CO<sub>2</sub> sink). Inversions indicate the net land CO<sub>2</sub> sink estimated by an ensemble of four atmospheric inversions. Dynamic Global Vegetation Models (DGVMs) indicate the mean net land CO<sub>2</sub> sink estimated by 17 dynamic global vegetation models driven by climate change, rising atmospheric CO<sub>2</sub>, land use change and nitrogen deposition change (for carbon-nitrogen models). The positive values indicate net CO<sub>2</sub> uptake from the atmosphere. (b) Normalised difference vegetation index (NDVI). The anomaly of global area-weighted NDVI observed by AVHRR and MODIS satellite sensors. AVHRR data are accessible during 1982–2016 and MODIS data are accessible during 2000–2018. (c) Near-infrared reflectance of vegetation (NIRv) and contiguous solar-induced chlorophyll fluorescence (CSIF). The standardised anomaly of area-weighted NIRv during 2001–2018 (Badgley et al., 2017) and CSIF during 2000–2018 (Zhang et al., 2018). (d) Gross primary production (GPP). The GPP from Cheng et al. (2017), DGVMs and MODIS GPP product (MOD17A3). GPP from Cheng et al. (2017) is based on an analytical model driven by climate change, rising atmospheric CO<sub>2</sub>, AVHRR leaf area index datasets and evapotranspiration datasets. GPP from DGVMs is the ensemble mean global GPP estimated by the same 17 DGVMs that provide the net land CO<sub>2</sub> sink estimates. Shaded area indicates 1-σ inter-model spread except for atmospheric inversions, whose ranges were used due to limited number of models. Further details on data sources and processing are available in the chapter data table (Table 5.SM.6).

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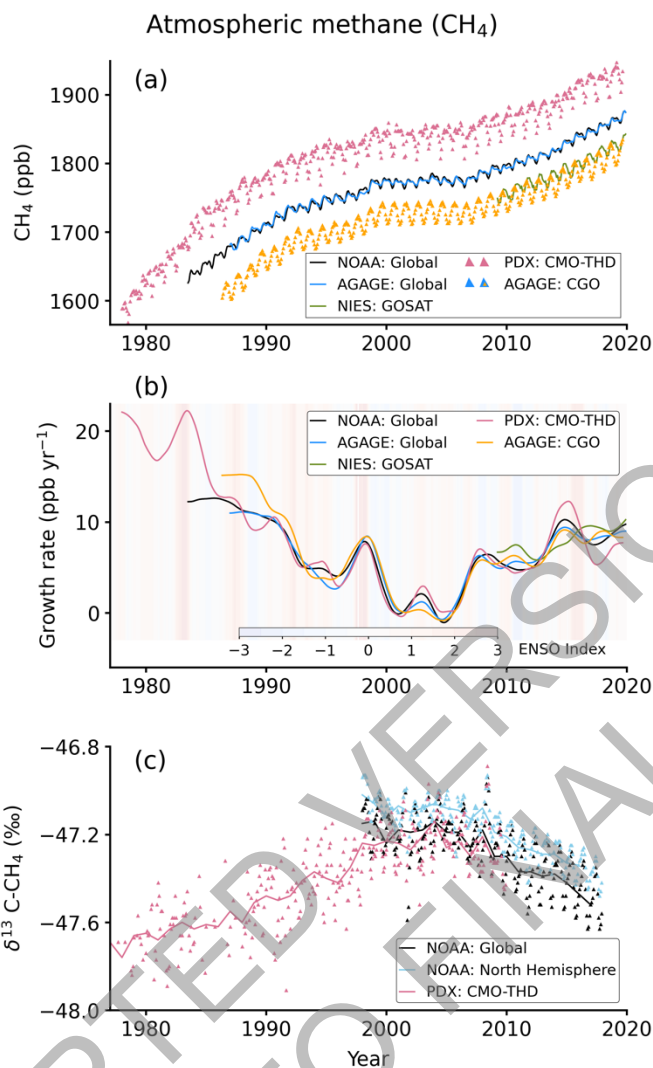


**Figure 5.11: Interannual variation in detrended anomalies of the net land CO<sub>2</sub> sink and land surface air temperature during 1980–2019.** Correlation coefficients between the net land CO<sub>2</sub> sink anomalies and temperature anomalies are show on the right bar plots. The net land CO<sub>2</sub> sink is estimated by four atmospheric inversions (blue) and fifteen Dynamic Global Vegetation Models (DGVMs) (green), respectively (Friedlingstein et al., 2020). Solid blue and green lines show model mean detrended anomalies of the net land CO<sub>2</sub> sink. The ensemble mean of DGVMs is bounded by the 1-σ inter-model spread in each large latitude band (North 30°N–90°N, Tropics 30°S–30°N, South 90°S–30°S) and the globe. The ensemble mean of atmospheric inversions is bounded by model spread. For each latitudinal band, the anomalies of the net land CO<sub>2</sub> sink and temperature (orange) were obtained by removing the long-term trend and seasonal cycle. A 12-month running mean was taken to reduce high-frequency noise. The bars in the right panels show correlation coefficients between the net land CO<sub>2</sub> sink anomalies and temperature anomalies for each region. Two asterisks indicate P<0.01, and one indicates P<0.05. Grey shaded area shows the intensity of El Niño southern oscillation (ENSO) as defined by the Niño 3.4 index. Two volcanic eruptions (El Chichón eruption and Pinatubo eruption) are indicated with light blue dashed lines. Temperature data are from the Climatic Research Unit (CRU), University of East Anglia (Harris et al., 2014). Anomalies were calculated following Patra et al. (2005), but using 12-month low-pass filter and detrended to obtain interannual variations. Further details on data sources and processing are available in the chapter data table (Table 5.SM.6).

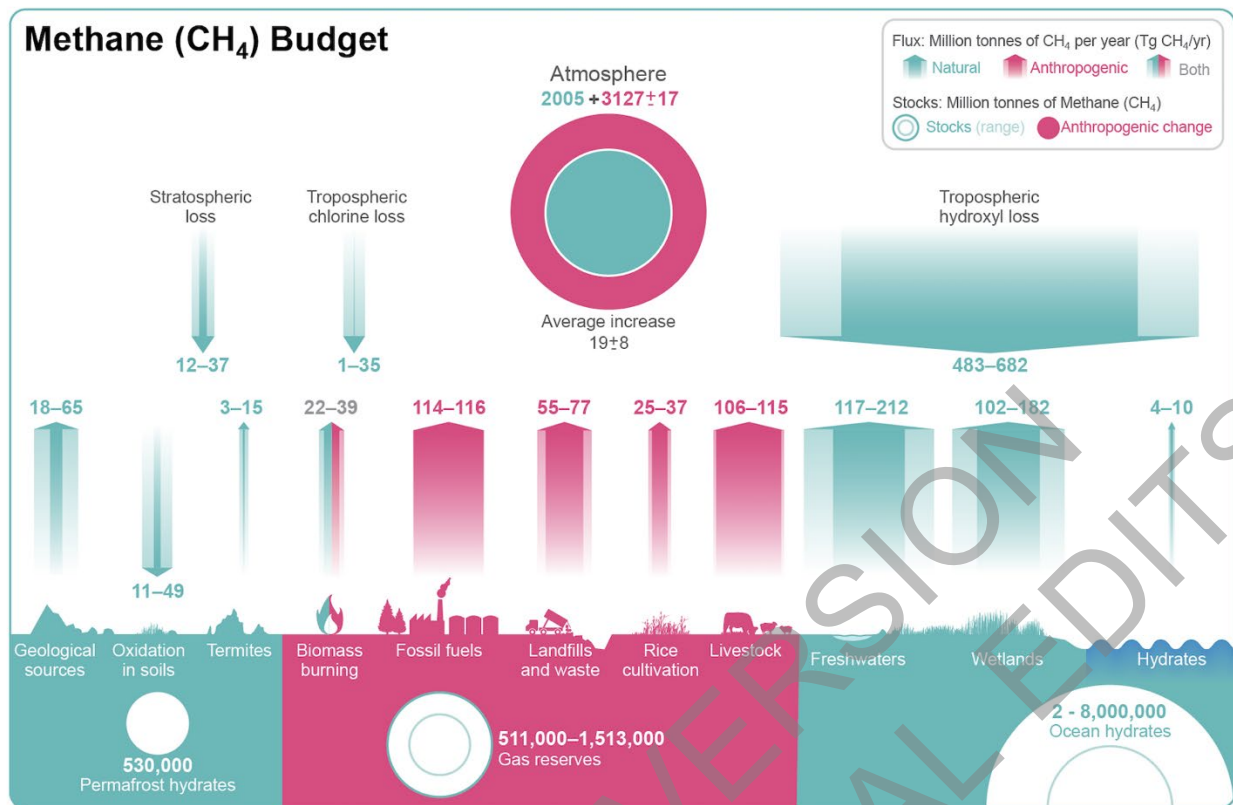
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**Figure 5.12: Global carbon (CO<sub>2</sub>) budget (2010–2019).** Yellow arrows represent annual carbon fluxes (in PgC yr<sup>-1</sup>) associated with the natural carbon cycle estimated for the time prior to the industrial era, around 1750. Pink arrows represent anthropogenic fluxes averaged over the period 2010–2019. The rate of carbon accumulation in the atmosphere is equal to net land-use change emissions, including land management (called LULUCF in the main text) plus fossil fuel emissions, minus land and ocean net sinks (plus a small budget imbalance, Table 5.1). Circles with yellow numbers represent pre-industrial carbon stocks in PgC. Circles with pink numbers represent anthropogenic changes to these stocks (cumulative anthropogenic fluxes) since 1750. Anthropogenic net fluxes are reproduced from Friedlingstein et al., (2020). The relative change of *gross photosynthesis* since pre-industrial times is based on 15 DGVMs used in Friedlingstein et al., (2020). The corresponding emissions by *Total respiration and fire* are those required to match the *net land flux*, exclusive of net land-use change emissions which are accounted for separately. The cumulative change of anthropogenic carbon in the terrestrial reservoir is the sum of carbon cumulatively lost by net land use change emissions, and net carbon accumulated since 1750 in response to environmental drivers (warming, rising CO<sub>2</sub>, nitrogen deposition). The adjusted gross natural ocean-atmosphere CO<sub>2</sub> flux was derived by rescaling the value in figure 1 of (Sarmiento and Gruber, 2002) of 70 PgC/yr by the revised estimate of the bomb <sup>14</sup>C inventory in the ocean. The original bomb <sup>14</sup>C inventory yielded an average global gas transfer velocity of 22 cm/hr; the revised estimate is 17cm/h leading to 17/22\*70=54. Dissolved organic carbon reservoir and fluxes from (Hansell et al., 2009). Dissolved inorganic carbon exchanges between surface and deep ocean, subduction and obduction from (Levy et al., 2013) Levy et al. 2013. Export production and flux from (Boyd et al., 2019). NPP and remineralisation in surface layer of the ocean from (Kwiatkowski et al., 2020; Séférian et al., 2020). Deep ocean reservoir from (Keppler et al., 2020). Note that the mass balance of the two ocean carbon stocks *surface ocean* and *intermediate and deep ocean* includes a yearly accumulation of anthropogenic carbon (not shown). Fossil fuel reserves are from (BGR, 2019); fossil fuel resources are 11,490 PgC for coal, 6,780 PgC for oil, and 365 PgC for natural gas. Permafrost region stores are from (Hugelius et al., 2014; Strauss et al., 2017; Mishra et al., 2021) (see also Box 5.1) and soil carbon stocks outside of permafrost region from Batjes, (2016); Jackson et al., (2017). Biomass stocks (range of seven estimates) are from Erb et al., (2018). Sources for the fluxes of the continuum land-to-ocean are provided in main text and adjusted within the ranges of the various assessment to balance the budget (section 5.2.1.5).



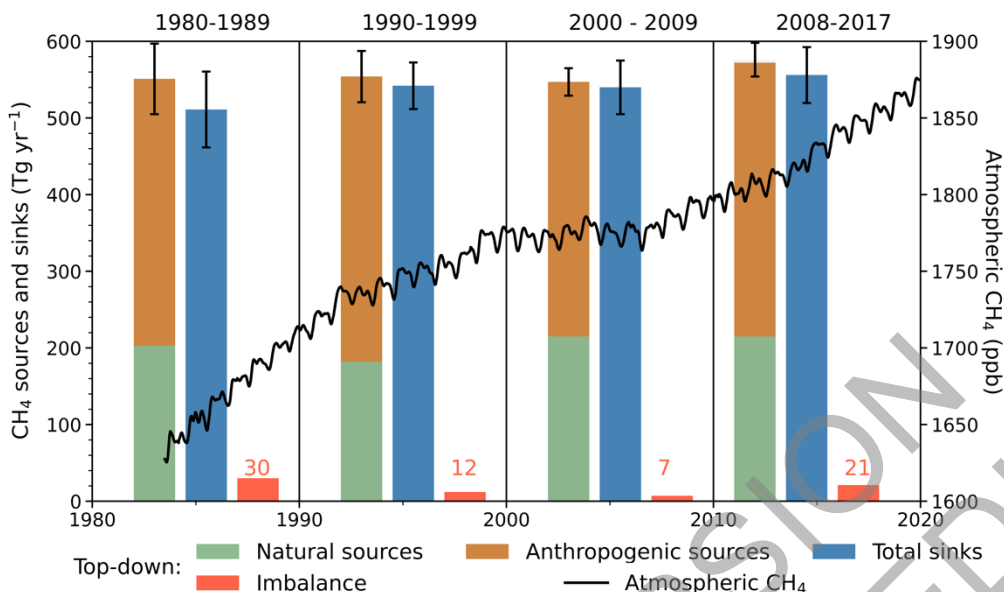
**Figure 5.13: Time series of CH<sub>4</sub> concentrations, growth rates and isotopic composition.** a) CH<sub>4</sub> concentrations, b) CH<sub>4</sub> growth rates, c)  $\delta^{13}\text{C}-\text{CH}_4$ . Data from selected site networks operated by NOAA (Dlugokencky et al., 2003), AGAGE (Prinn et al., 2018) and PDX (Portland State University) (Rice et al., 2016). To maintain clarity, data from many other measurement networks are not included here, and all measurements are shown in WMO X2004ACH<sub>4</sub> global calibration standard. Global mean values of XCH<sub>4</sub> (total-column), retrieved from radiation spectra measured by the greenhouse gases observation satellite (GOSAT) are shown in panels (a) and (b). Cape Grim Observatory (CGO; 41°S, 145°E) and Trinidad Head (THD; 41°N, 124°W) data are taken from the AGAGE network, NOAA global and northern hemispheric (NH) means for  $\delta^{13}\text{C}$  are calculated from 10 and 6 sites, respectively. The PDX data adjusted to NH (period: 1977–2000) are merged with THD (period: 2001–2019) for CH<sub>4</sub> concentration and growth rate analysis, and PDX and NOAA NH means of  $\delta^{13}\text{C}$  data are used for joint interpretation of long-term trends analysis. The multivariate ENSO index (MEI) is shown in panel (b). Further details on data sources and processing are available in the chapter data table (Table 5.SM.6).



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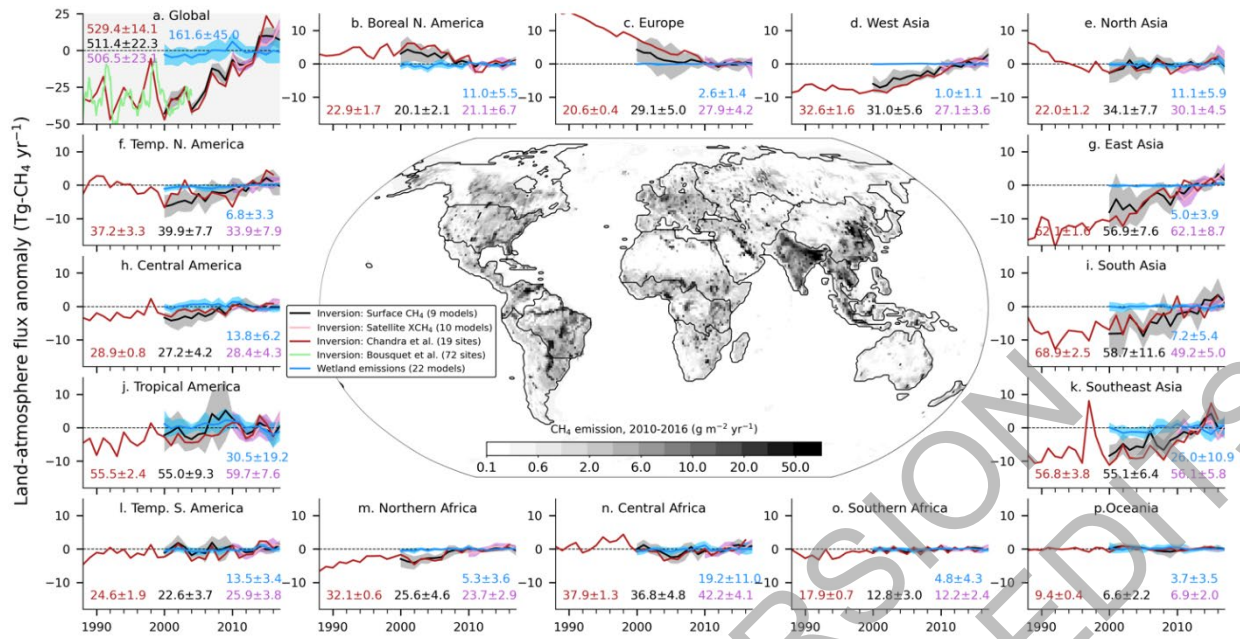
**Figure 5.14: Global methane (CH<sub>4</sub>) budget (2008–2017).** Values and data sources as in Table 5.2 (in TgCH<sub>4</sub>). The atmospheric stock is calculated from mean CH<sub>4</sub> concentration, multiplying a factor of  $2.75 \pm 0.015$  Tg ppb<sup>-1</sup>, which accounts for the uncertainties in global mean CH<sub>4</sub> (Chandra et al., 2021). Further details on data sources and processing are available in the chapter data table (Table 5.SM.6).

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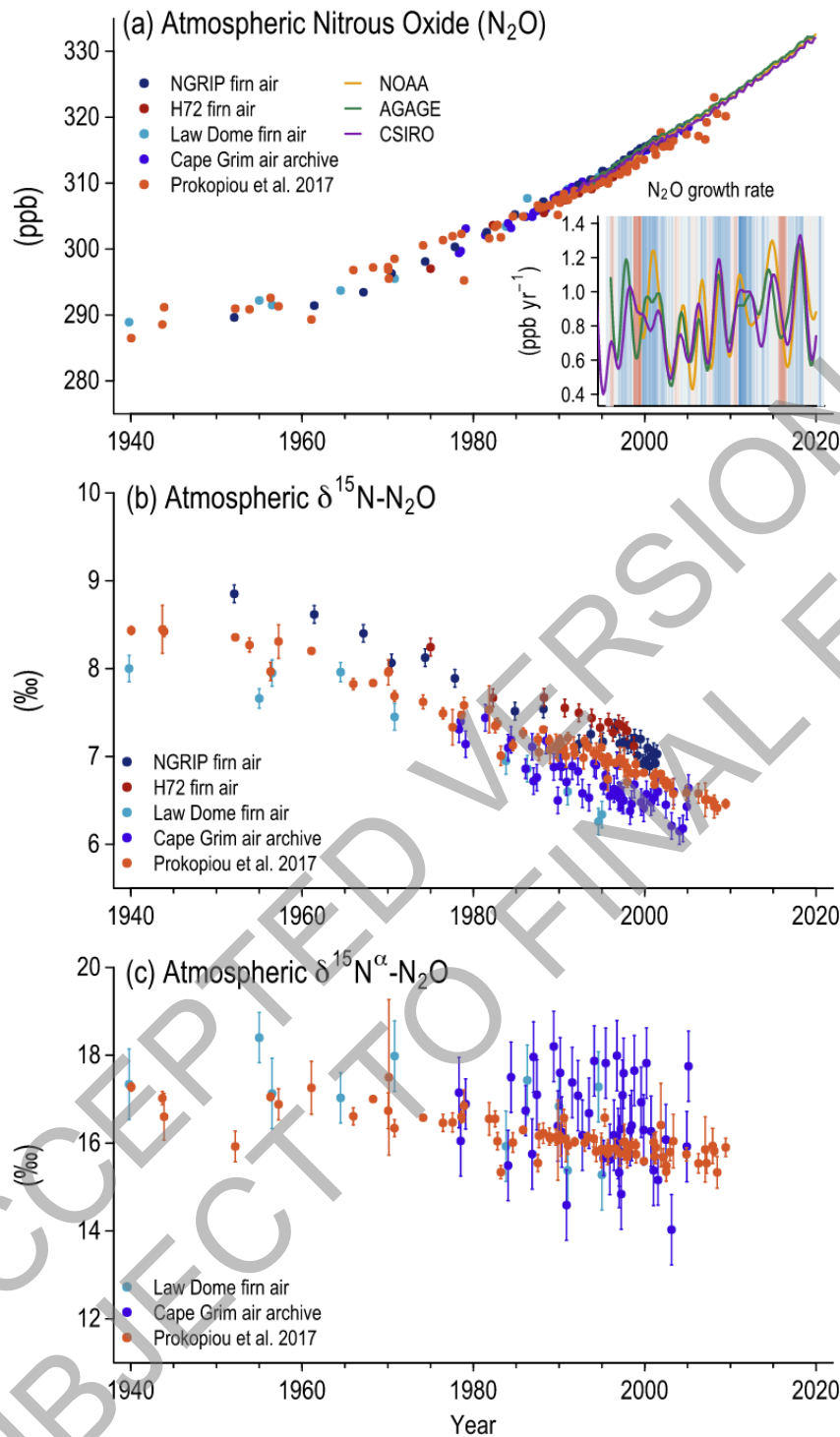


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**Cross-Chapter Box 5.2, Figure 1: Methane sources and sinks for four decades from atmospheric inversions with the budget imbalance** (source-sink; red bars) (plotted on the left y-axis). Top-down analysis from (Kirschke et al., 2013; Saunois et al., 2020), The global CH<sub>4</sub> concentration seen in the black line (plotted on the right y-axis), representing NOAA observed global monthly mean atmospheric CH<sub>4</sub> in dry-air mole fractions for 1983–2019 (Chapter 2, Annex V). Natural sources include emissions from natural wetlands, lakes and rivers, geological sources, wild animals, termites, wildfires, permafrost soils, and oceans. Anthropogenic sources include emissions from enteric fermentation and manure, landfills, waste and wastewater, rice cultivation, coal mining, oil and gas industry, biomass and biofuel burning. The top-down total sink is determined from global mass balance includes chemical losses due to reactions with hydroxyl (OH), atomic chlorine (Cl), and excited atomic oxygen (O<sup>1</sup>D), and oxidation by bacteria in aerobic soils (Table 5.2). Further details on data sources and processing are available in the chapter data table (Table 5.SM.6).



**Cross-Chapter Box 5.2, Figure 2: Anomalies in global and regional methane (CH<sub>4</sub>) emissions for 1988–2017.** Map in the centre shows mean CH<sub>4</sub> emission for 2010–2016. Multi-model mean (line) and 1- $\sigma$  standard deviations (shaded) for 2000–2017 are shown for 9 surface CH<sub>4</sub> and 10 satellite XCH<sub>4</sub> inversions, and 22 wetland models or model variants that participated in GCP-CH<sub>4</sub> budget assessment (Saunio et al., 2020). The results for the period before 2000 are available from two inversions, 1) using 19 sites (Chandra et al., 2021; also used for the 2010–2016 mean emission map) and for global totals (Bousquet et al., 2006). The long-term mean values for 2010–2016 (common for all GCP-CH<sub>4</sub> inversions), as indicated within each panel separately, is subtracted from the annual-mean time series for the calculation of anomalies for each region. Further details on data sources and processing are available in the chapter data table (Table 5.SM.6).

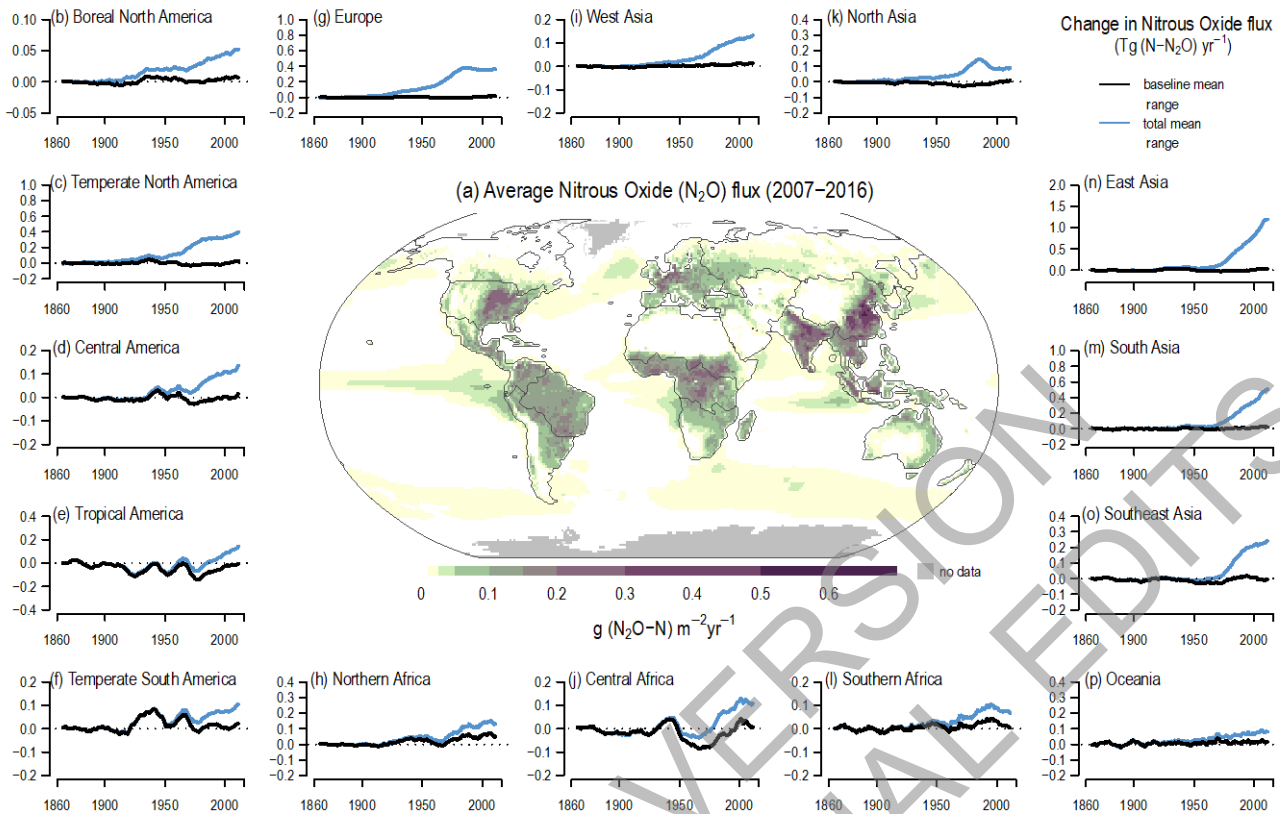


**Figure 5.15: Changes in atmospheric nitrous oxide (N<sub>2</sub>O) and its isotopic composition since 1940. (a)**

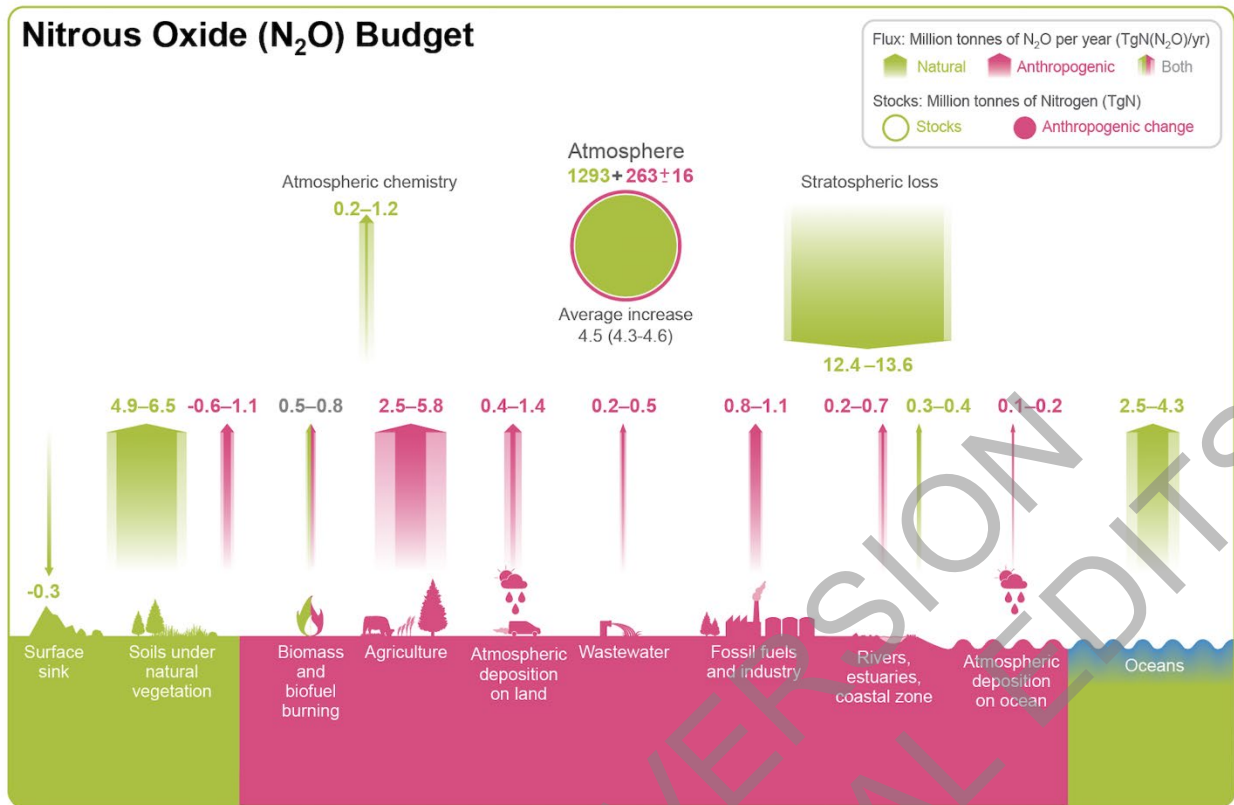
Atmospheric N<sub>2</sub>O abundance (parts per billion, ppb) and growth rate (ppb yr<sup>-1</sup>), (b) δ<sup>15</sup>N of atmospheric N<sub>2</sub>O, and (c) alpha-site <sup>15</sup>N-N<sub>2</sub>O. Estimate are based on direct atmospheric measurements in the AGAGE, CSIRO, and NOAA networks (Prinn et al., 2000, 2018; Francey et al., 2003; Hall et al., 2007; Elkins et al., 2018), archived air samples from Cape Grim, Australia (Park et al., 2012), and firn air from NGRIP Greenland and H72 Antarctica (Ishijima et al., 2007), Law Dome Antarctica (Park et al., 2012), as well as a collection of firn ice samples from Greenland (Prokopiou et al., 2017, 2018). Shading in (a) is based on the multivariate ENSO index, with red indicating El Niño conditions (Wolter and Timlin, 1998). Further details on data sources and processing are available in the chapter data table (Table 5.SM.6).



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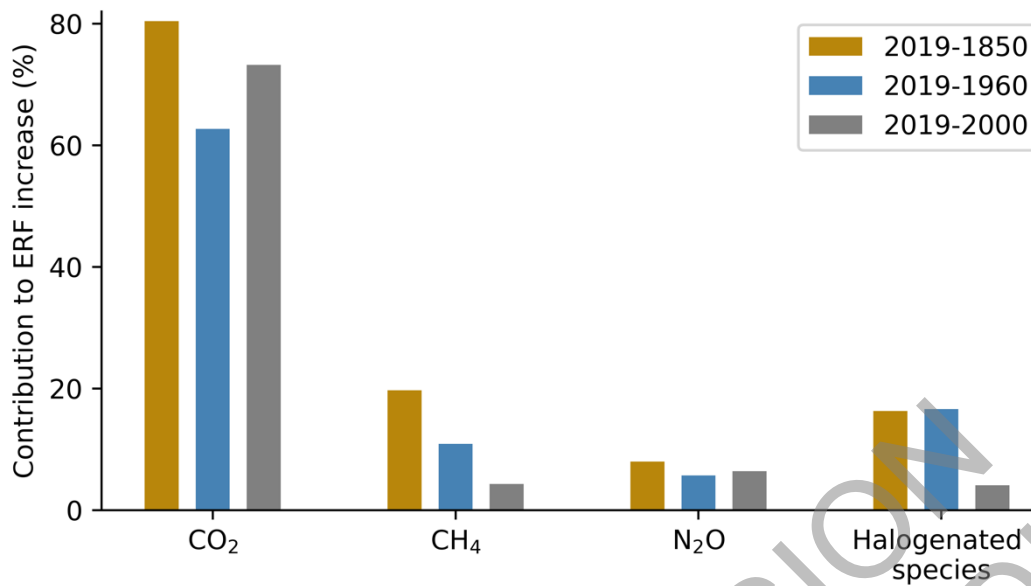
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**Figure 5.16: Decadal mean nitrous oxide ( $\text{N}_2\text{O}$ ) emissions for 2007–2016 and its change since 1850 based on process-model projections.** The total effect including that from anthropogenic nitrogen additions (atmospheric deposition, manure addition, fertiliser use and land-use) is evaluated against the background flux driven by changes in atmospheric  $\text{CO}_2$  concentration, and climate change. Fluxes are derived from the  $\text{N}_2\text{O}$  Model Intercomparison Project ensemble of terrestrial biosphere models (Tian et al., 2019) and three ocean biogeochemical models (Landolfi et al., 2017; Battaglia and Joos, 2018a; Buitenhuis et al., 2018). Further details on data sources and processing are available in the chapter data table (Table 5.SM.6).



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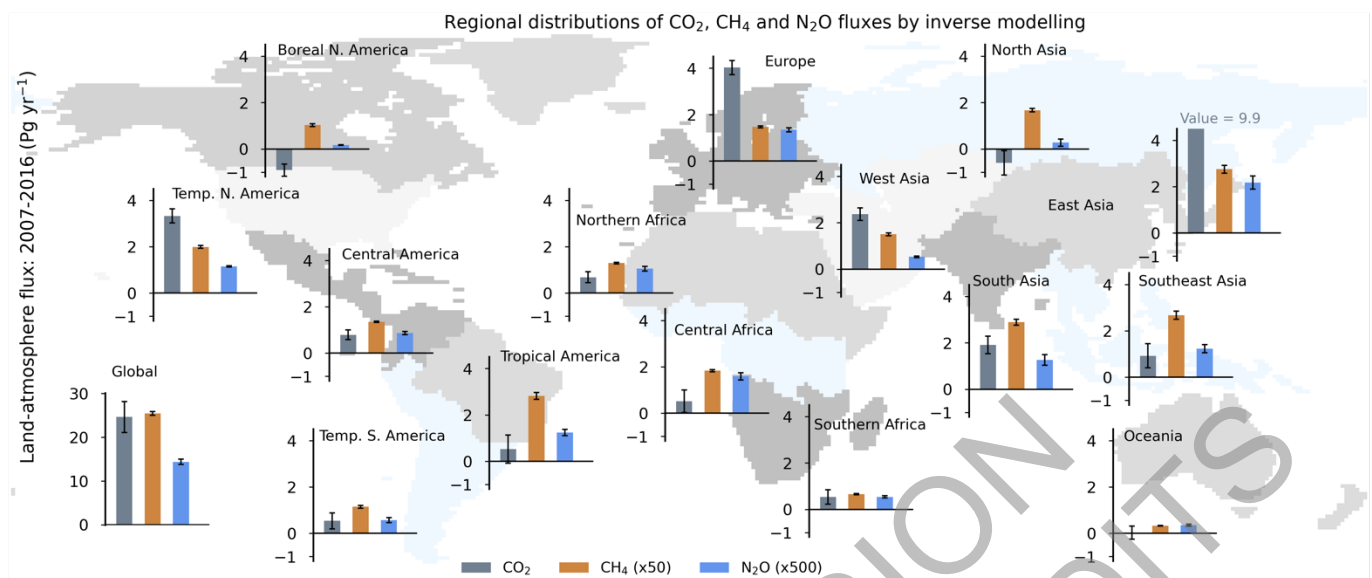
**Figure 5.17: Global nitrous oxide (N<sub>2</sub>O) budget (2007–2016).** Values and data sources as in Table 5.3. The atmospheric stock is calculated from mean N<sub>2</sub>O concentration, multiplying a factor of  $4.79 \pm 0.05 \text{ Tg ppb}^{-1}$  (Prather et al., 2012). Pool sizes for the other reservoirs are largely unknown. Further details on data sources and processing are available in the chapter data table (Table 5.SM.6).



**Figure 5.18: Contributions of carbon dioxide (CO<sub>2</sub>), methane (CH<sub>4</sub>), nitrous oxide (N<sub>2</sub>O) and halogenated species to the total effective radiative forcing (ERF) increase since 1850 and 1960, and for 2000 to 2009.** ERF data are taken from Annex 5 (based on calculations from Chapter 7). Note that the sum of the ERFs exceeds 100% because there are negative ERFs due to aerosols and clouds. Further details on data sources and processing are available in the chapter data table (Table 5.SM.6).

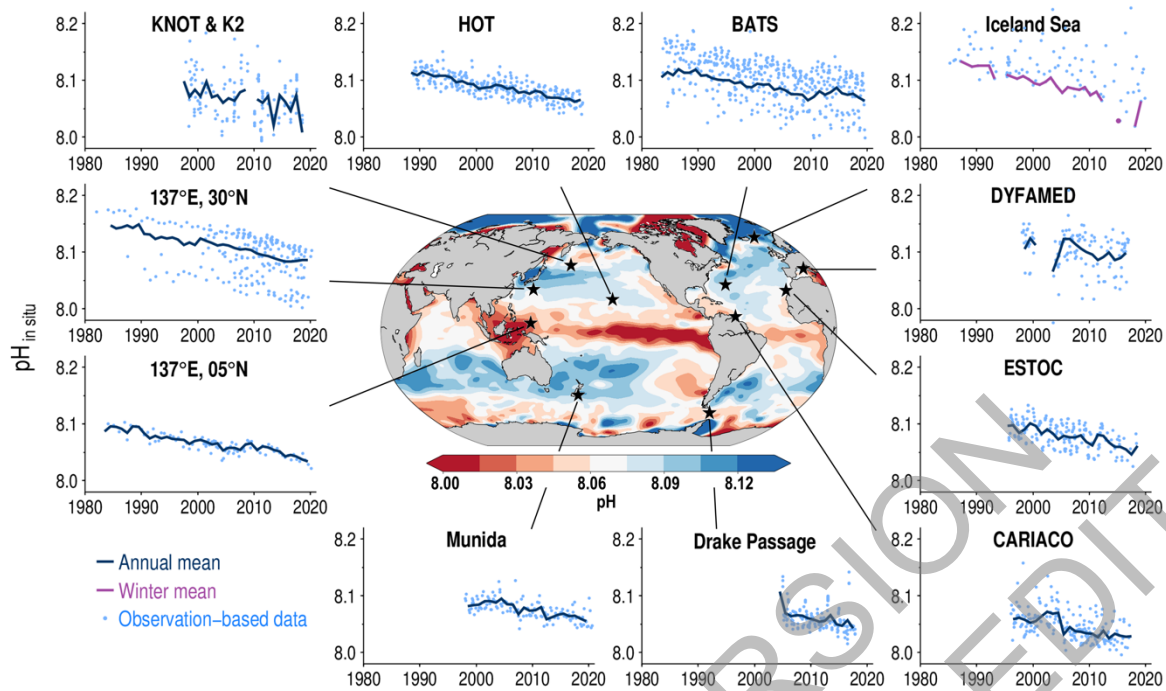
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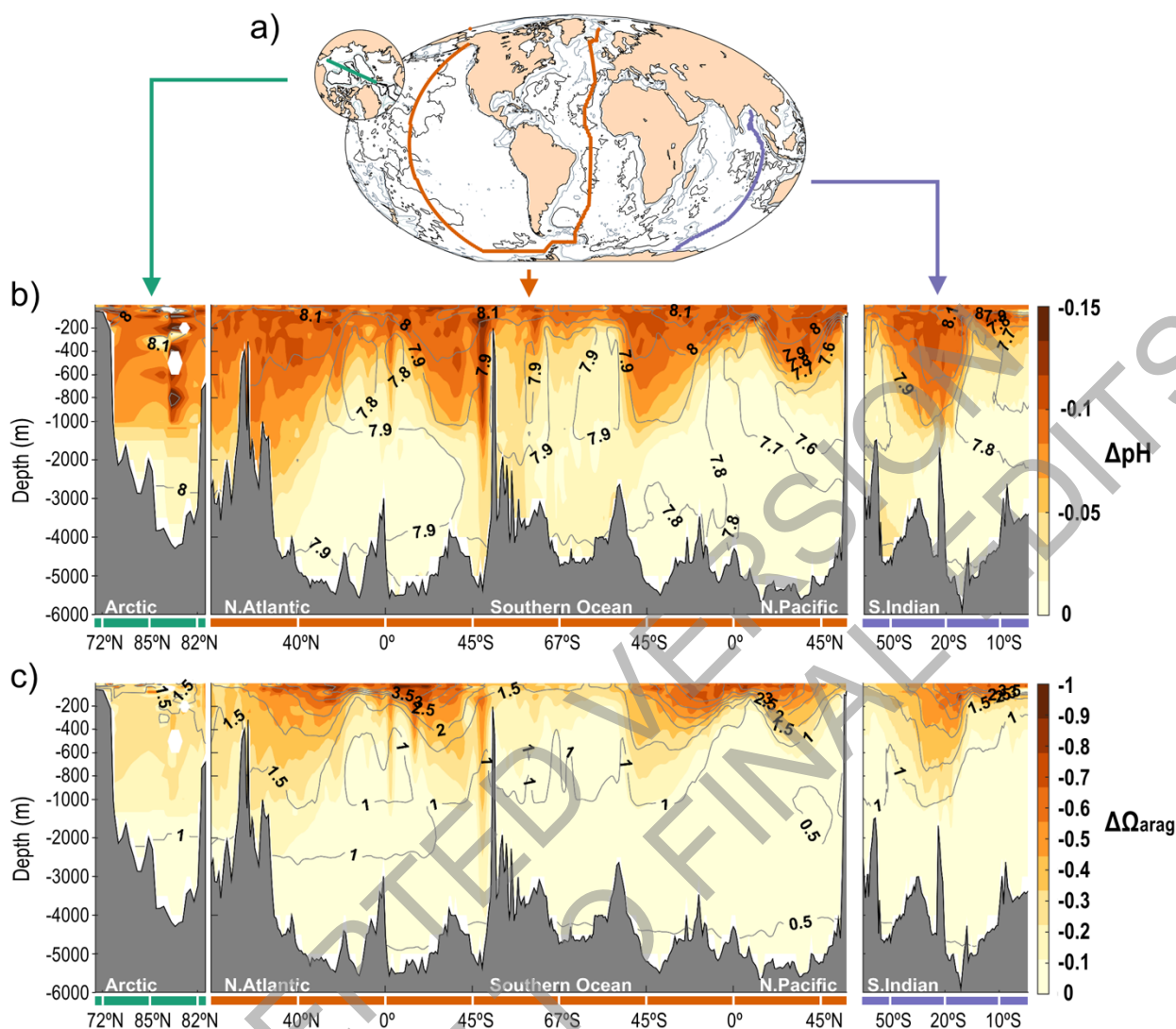


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**Figure 5.19: Regional distributions of net fluxes of carbon dioxide (CO<sub>2</sub>), methane (CH<sub>4</sub>), nitrous oxide (N<sub>2</sub>O) on the Earth's surface.** The region divisions, shown as the shaded map, are made based on ecoclimatic characteristics of the land. The fluxes include those from anthropogenic activities and natural causes that result from responses to anthropogenic greenhouse gases and climate forcing (feedbacks) as in the three budgets shown in Sections 5.2.1.5, 5.2.2.5, and 5.2.3.5. The CH<sub>4</sub> and N<sub>2</sub>O emissions are weighted by arbitrary factors of 50 and 500, respectively, for depiction by common y-axes. Fluxes are shown as the mean of the inverse models as available from (Thompson et al., 2019; Friedlingstein et al., 2020; Saunio et al., 2020). Further details on data sources and processing are available in the chapter data table (Table 5.SM.6).



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3 **Figure 5.20: Multi-decadal trends of pH (Total Scale) in surface layer at various sites of the oceans and a global**  
4 **distribution of annual mean pH adjusted to the year 2000.** Time-series data of pH are from Dore et  
5 al., 2009; Olafsson et al., 2009; González-Dávila et al., 2010; Bates et al., 2014b; Takahashi et al., 2014;  
6 Wakita et al., 2017; Merlivat et al., 2018; Ono et al., 2019; and Bates and Johnson, 2020. Global  
7 distribution of annual mean pH have been evaluated from data of surface ocean  $p\text{CO}_2$  measurements  
8 (Bakker et al., 2016; Jiang et al., 2019). Acronyms in panels: KNOT and K2 - Western Pacific subarctic  
9 gyre time-series; HOT - Hawaii Ocean Time-series; BATS - Bermuda Atlantic Time-series Study;  
10 DYFAMED - Dynamics of Atmospheric Fluxes in the Mediterranean Sea; ESTOC - European Station for  
11 Time-series in the Ocean Canary Islands; CARIACO - Carbon Retention in a Colored Ocean Time-series.  
12 Further details on data sources and processing are available in the chapter data table (Table 5.SM.6).  
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2 **Figure 5.21: Spread of ocean acidification from the surface into the interior of ocean since pre-industrial times.**

3 (a) map showing the three transects used to create the cross sections shown in (b) and (c); vertical  
4 sections of the changes in (b) pH and (c) saturation state of aragonite ( $\Omega_{\text{arag}}$ ) between 1800–2002 due to  
5 anthropogenic  $\text{CO}_2$  invasion (colour). Contour lines are their contemporary values in 2002. The red  
6 transect begins in the Nordic Seas and then follows the GO-SHIP lines A16 southward in the Atlantic  
7 Ocean, SR04 and S04P westward in the Southern Ocean, and P16 northward in the Pacific Ocean. The  
8 purple line follows the GO-SHIP line I09 southward in the Indian Ocean. The green line on the smaller  
9 inset crosses the Arctic Ocean from the Bering Strait to North Pole along  $175^\circ\text{W}$  and from the North Pole  
10 to the Fram Strait along  $5^\circ\text{E}$  (Lauvset et al., 2020). Further details on data sources and processing are  
11 available in the chapter data table (Table 5.SM.6).  
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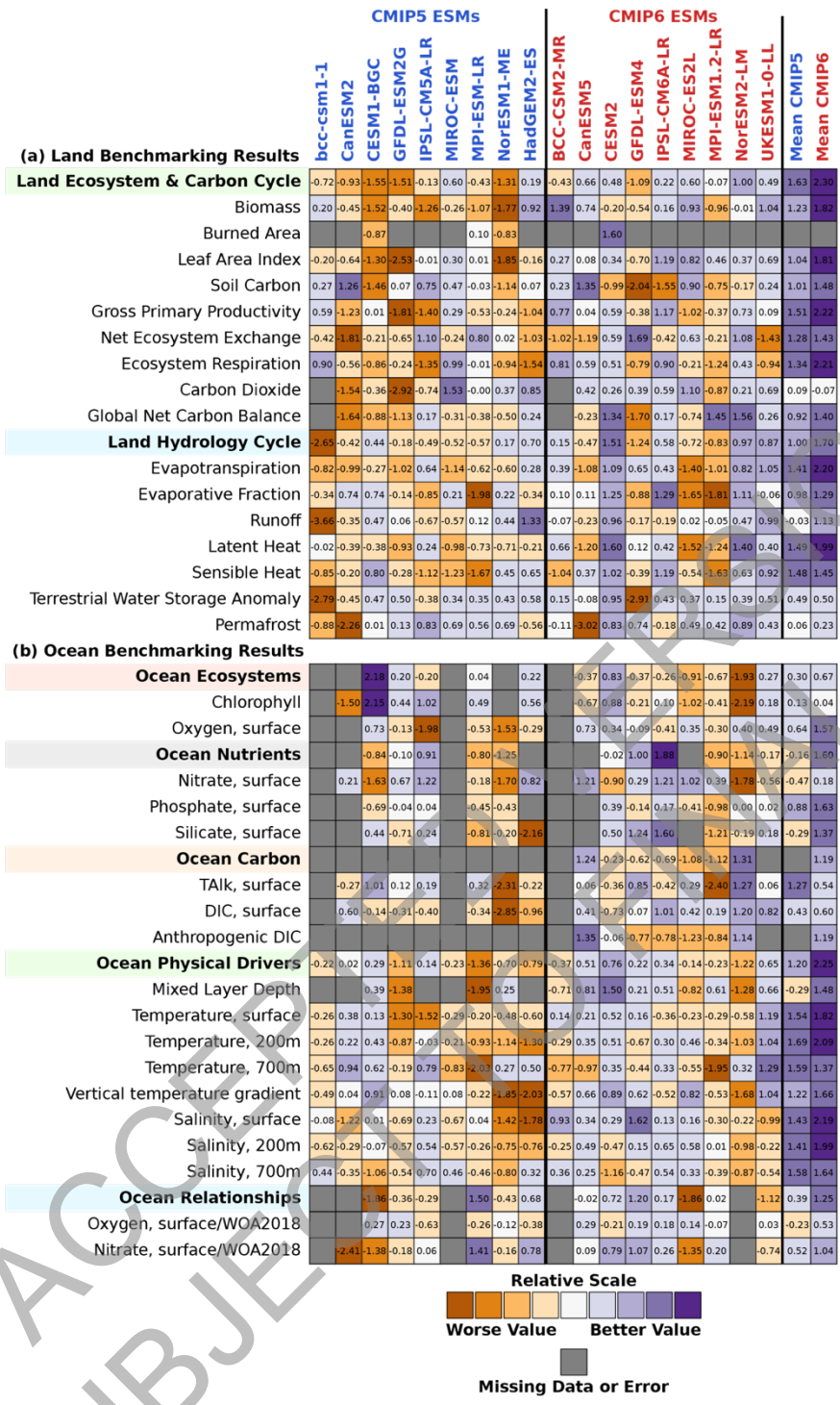
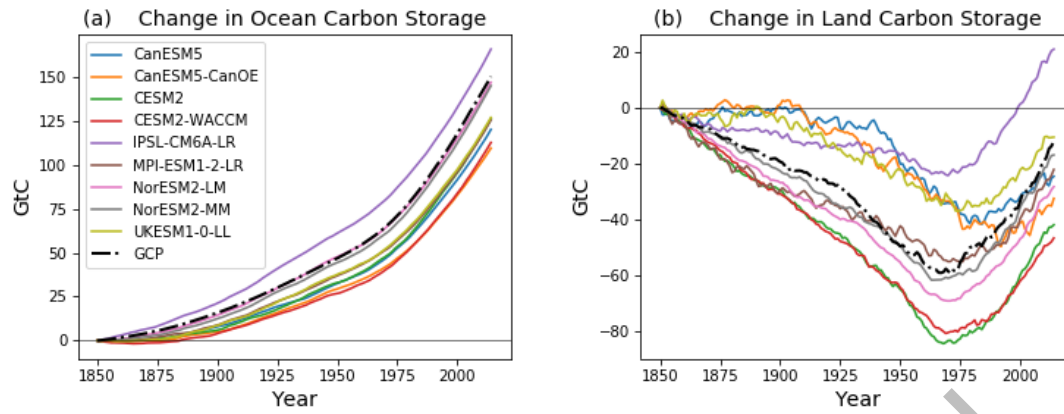


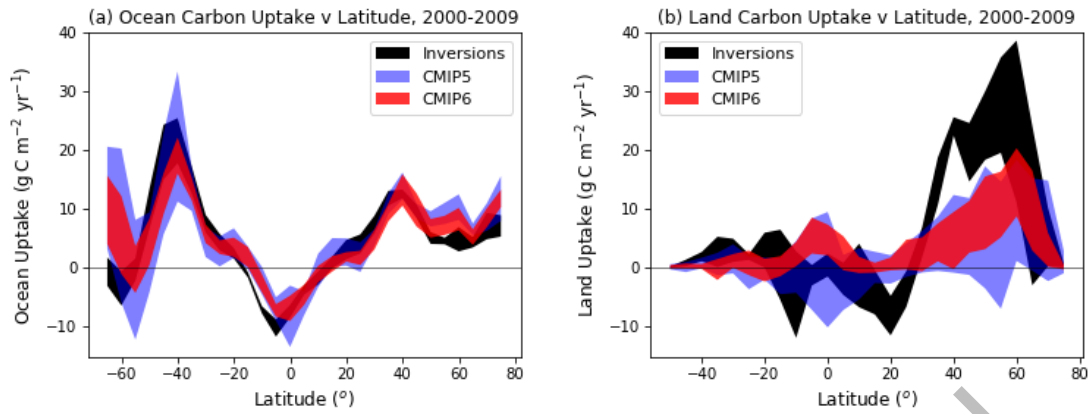
Figure 5.22: Overview scores for CMIP5 (left hand side of table) and CMIP6 (right hand side of table) Earth system models (ESMs), for multiple benchmarks against different datasets. (a) Benchmarking of ESM land models, (b) benchmarking of ocean models. Scores are relative to other models within each benchmark row, with positive scores indicating a better agreement with observations. Models included are only those from institutions that participated in both CMIP5 and CMIP6 carbon cycle experiments, in order to trace changes from one ensemble to the next. CMIP5 models are labels in yellow and CMIP6 in green, with the multi-model ensemble means labelled in white. Further details on data sources and processing are available in the chapter data table (Table 5.SM.6).

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3 **Figure 5.23: CMIP6 Earth system model (ESM) concentration-driven historical simulations for 1850 to 2014,**  
4 **compared to observation-based estimates from the global carbon project (GCP). Panel (a)**  
5 **cumulative ocean carbon uptake from 1850 (PgC); (b) cumulative land carbon uptake from 1850 (PgC).**  
6 **Only models that simulate both land and ocean carbon fluxes are shown here. Further details on data**  
7 **sources and processing are available in the chapter data table (Table 5.SM.6).**  
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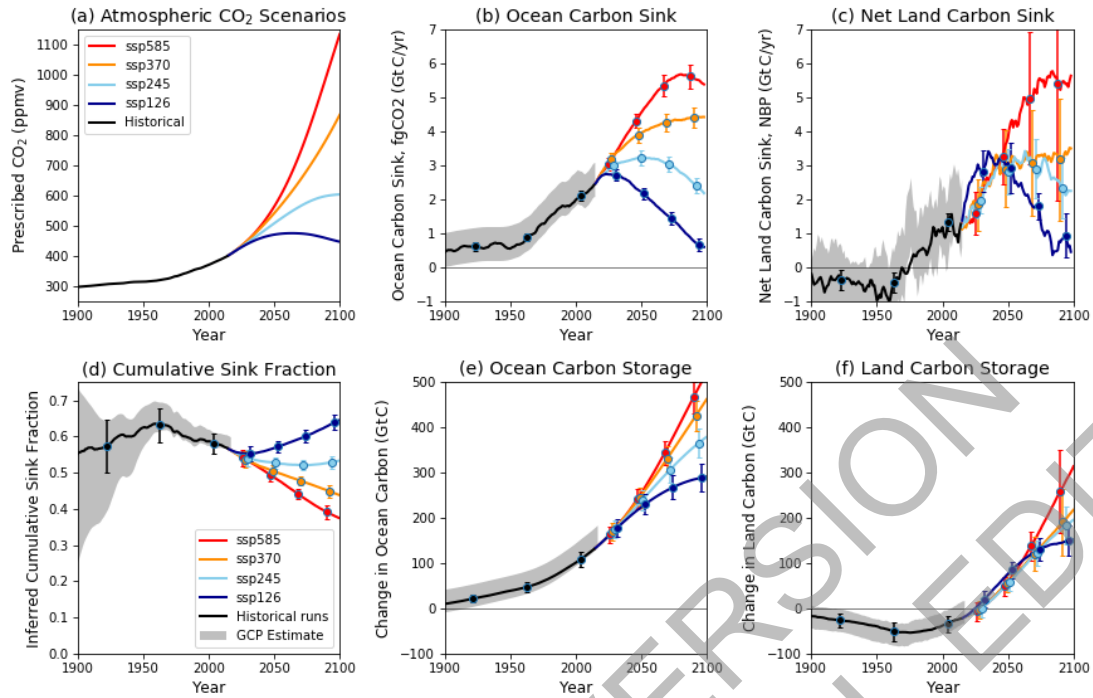




**Figure 5.24: Comparison of modelled zonal distribution of contemporary carbon sinks against atmospheric inversion estimates for 2000–2009, (a) ocean carbon uptake; (b) net land uptake.** Latitude runs from 90°S (i.e. –90°N) to 90°N. Positive uptake represents a carbon sink to ocean/land while negative uptake represents a carbon source. The land uptake is taken as Net Biome Productivity (NBP) and so includes net land-use change emissions. The bands show the mean  $\pm 1$  standard deviation across the available inversions (black bands, 3 models), CMIP5 Earth system models (ESMs) (blue bands, 12 models for the ocean, 12 models for the land), and CMIP6 ESMs (red bands, 11 models for ocean, 10 models for land). Further details on data sources and processing are available in the chapter data table (Table 5.SM.6).

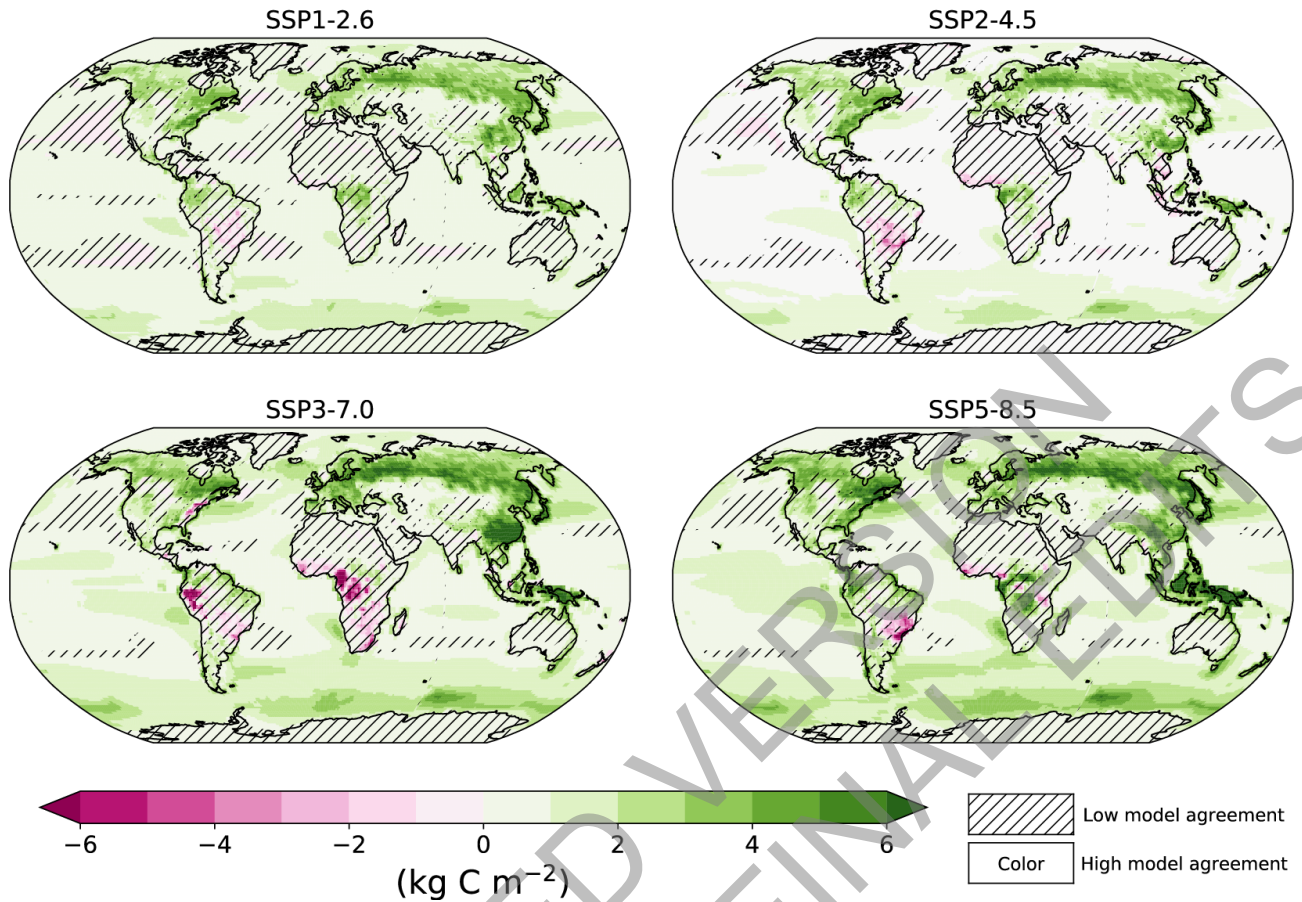
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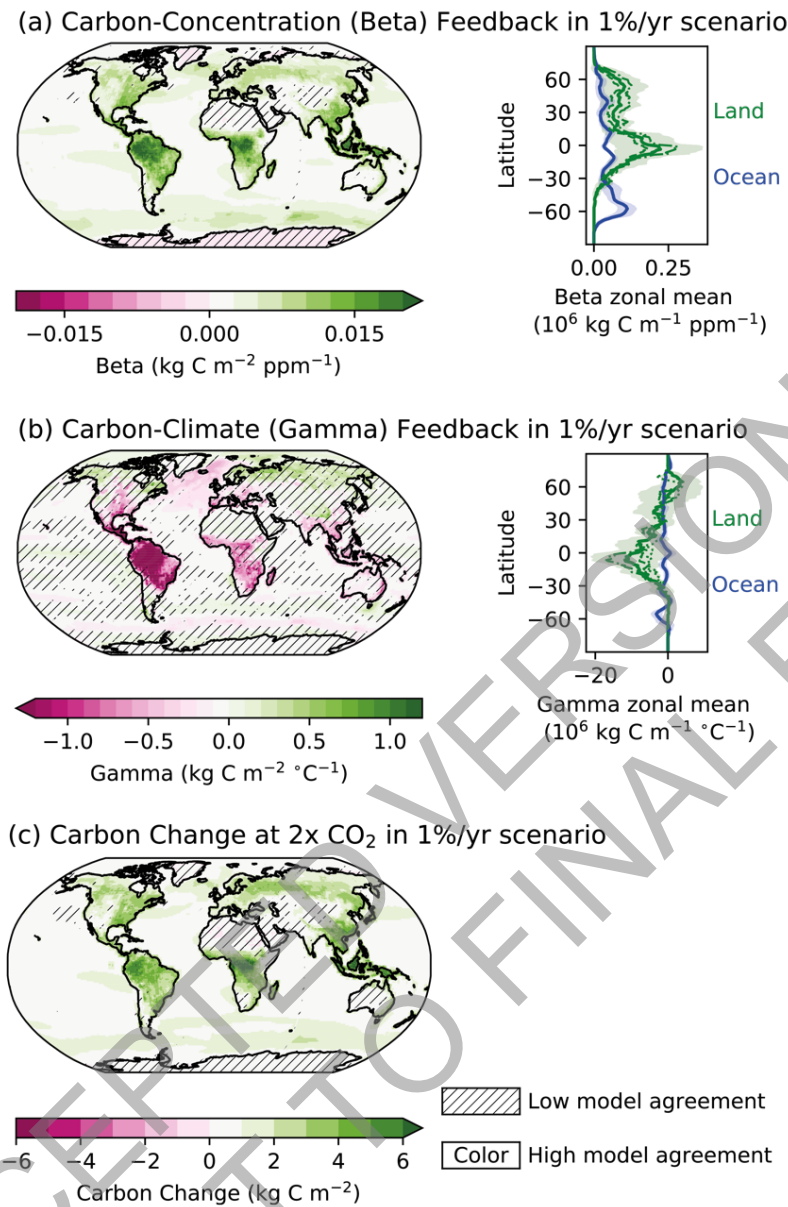
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 2 **Figure 5.25: Modelled evolution of the global land and ocean carbon sinks for 1850 to 2100 in concentration-**  
 3 **driven CMIP6 Earth system model (ESM) scenario runs** (SSP1–2.6: blue; SSP2–4.5: skyblue; SSP3–  
 4 7.0: yellow; SSP5–8.5: red): (a) prescribed atmospheric CO<sub>2</sub> concentrations; (b) 5-year running mean  
 5 ocean carbon sink (GtC yr<sup>-1</sup>); (c) 5-year running mean net land carbon sink (GtC yr<sup>-1</sup>); (d) inferred  
 6 cumulative sink fraction of emissions from 1850; (e) change in ocean carbon storage from 1850 (GtC); (f)  
 7 change in land carbon storage from 1850 (GtC). Thick lines represent the ensemble mean of the listed  
 8 ESM runs, and the error bars represents ± one standard deviation about that mean. The grey wedges  
 9 represent estimates from the global carbon project (GCP), assuming uncertainties in the annual mean  
 10 ocean and net land carbon sinks of 0.5 GtC yr<sup>-1</sup> and 1 GtC yr<sup>-1</sup> respectively, and uncertainties in the  
 11 changes in carbon stores (ocean, land and cumulative total emissions) of 25 GtC. The net land carbon  
 12 sink is taken as net biome productivity (NBP) and so includes any modelled net land-use change  
 13 emissions. Further details on data sources and processing are available in the chapter data table (Table  
 14 5.SM.6).

Change in carbon from 2015 to 2100 under SSP scenarios



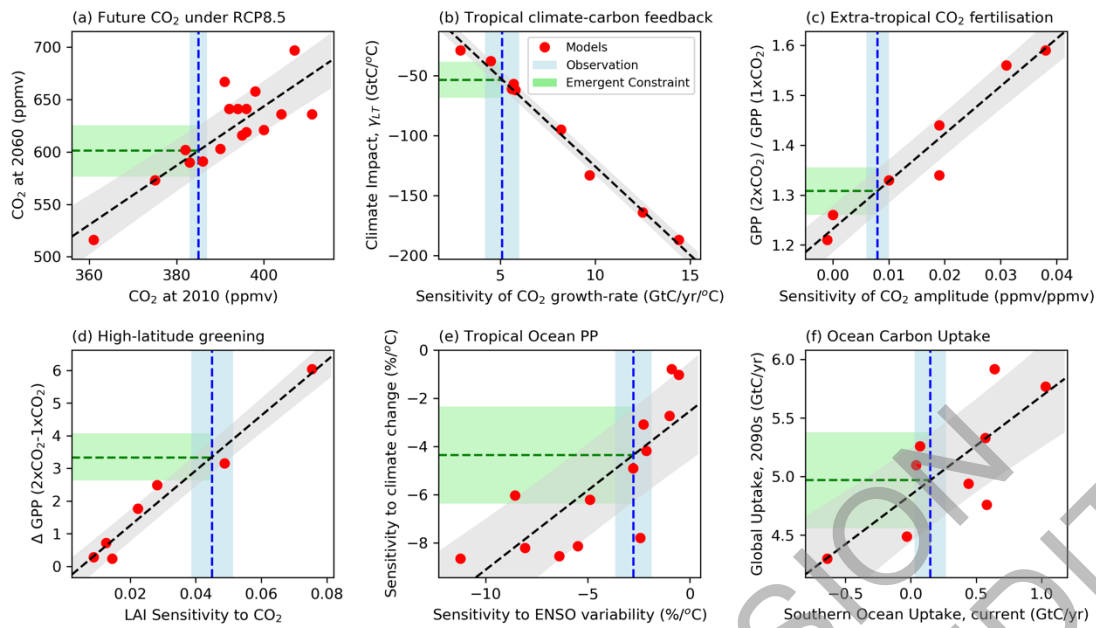
**Figure 5.26: Maps of net carbon changes under four Shared Socioeconomic Pathway (SSP) scenarios, as evaluated from nine CMIP6 Earth system models.** Uncertainty is represented using the simple approach (see Cross-Chapter Box Atlas.1 for more information): No overlay indicates regions with high model agreement, where  $\geq 80\%$  of models agree with the ensemble mean on the sign of change; diagonal lines indicate regions with low model agreement, where  $< 80\%$  of models agree with the ensemble mean on the sign of change. On land, this is calculated as the time integral of NBP, for the ocean it is the time-integral of air-sea  $\text{CO}_2$  gas flux anomalies relative to the pre-industrial. Further details on data sources and processing are available in the chapter data table (Table 5.SM.6).

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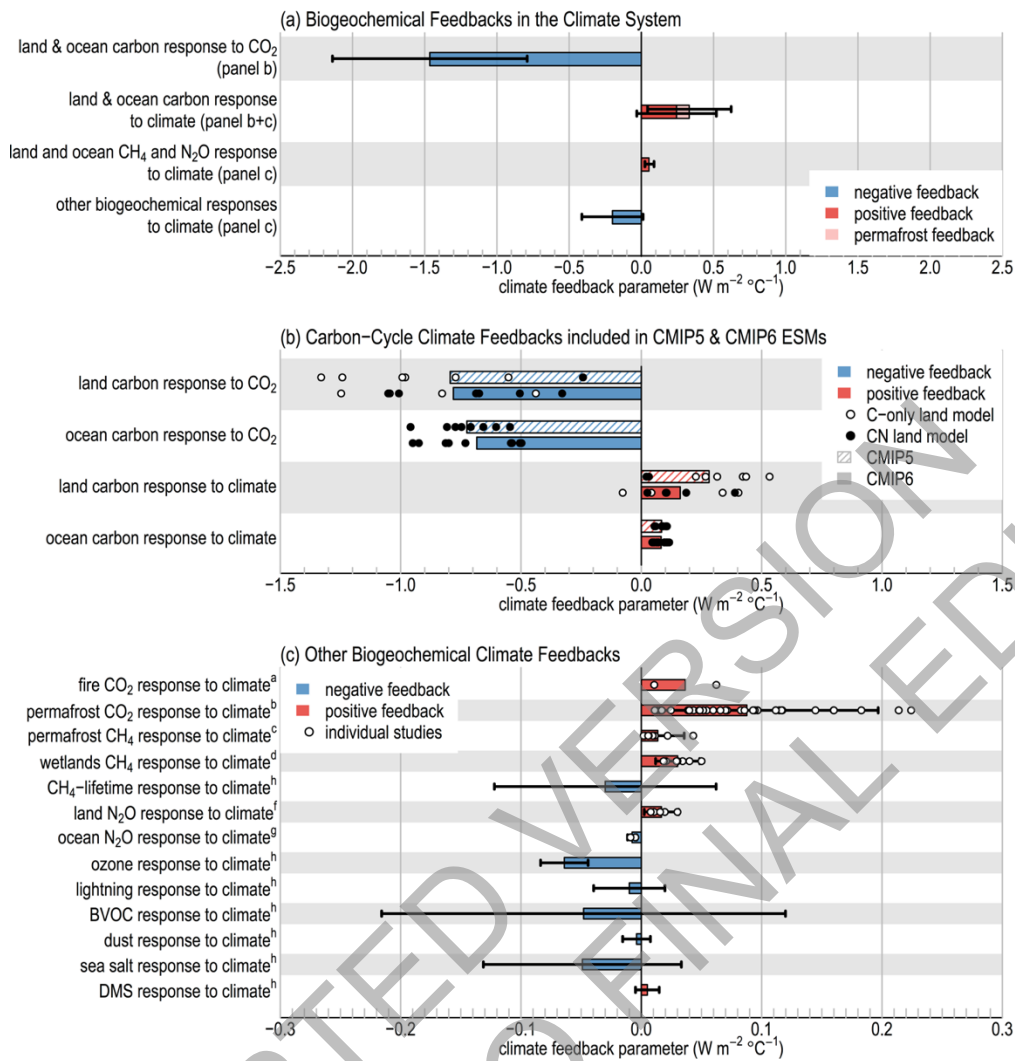
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**Figure 5.27: Maps of carbon-concentration and carbon-climate feedback terms, as well as net carbon changes under the idealised 1% per year CO<sub>2</sub> scenario, as evaluated from CMIP6 Earth system models (ESMs).** Shown are the model means from nine CMIP6 ESMs. Uncertainty is represented using the simple approach (see Cross-Chapter Box Atlas.1 for more information): No overlay indicates regions with high model agreement, where  $\geq 80\%$  of models agree with the ensemble mean on the sign of change; diagonal lines indicate regions with low model agreement, where  $< 80\%$  of models agree with the ensemble mean on the sign of change. Also shown are zonal-mean latitude profiles of land (green) and ocean (blue) feedbacks. On the land, the zonal mean feedback for the mean of the ensemble of models that include nitrogen is shown as dashed lines, and carbon-only models as dash-dotted lines, and the carbon-climate feedback from one permafrost-carbon enabled ESM is shown as a dotted line carbon changes are calculated as the difference between carbon stocks at different times on land and for the ocean as the time integral of atmosphere-ocean CO<sub>2</sub> flux anomalies relative to the pre-industrial. The denominator for gamma here is the global mean surface air temperature. Further details on data sources and processing are available in the chapter data table (Table 5.SM.6).



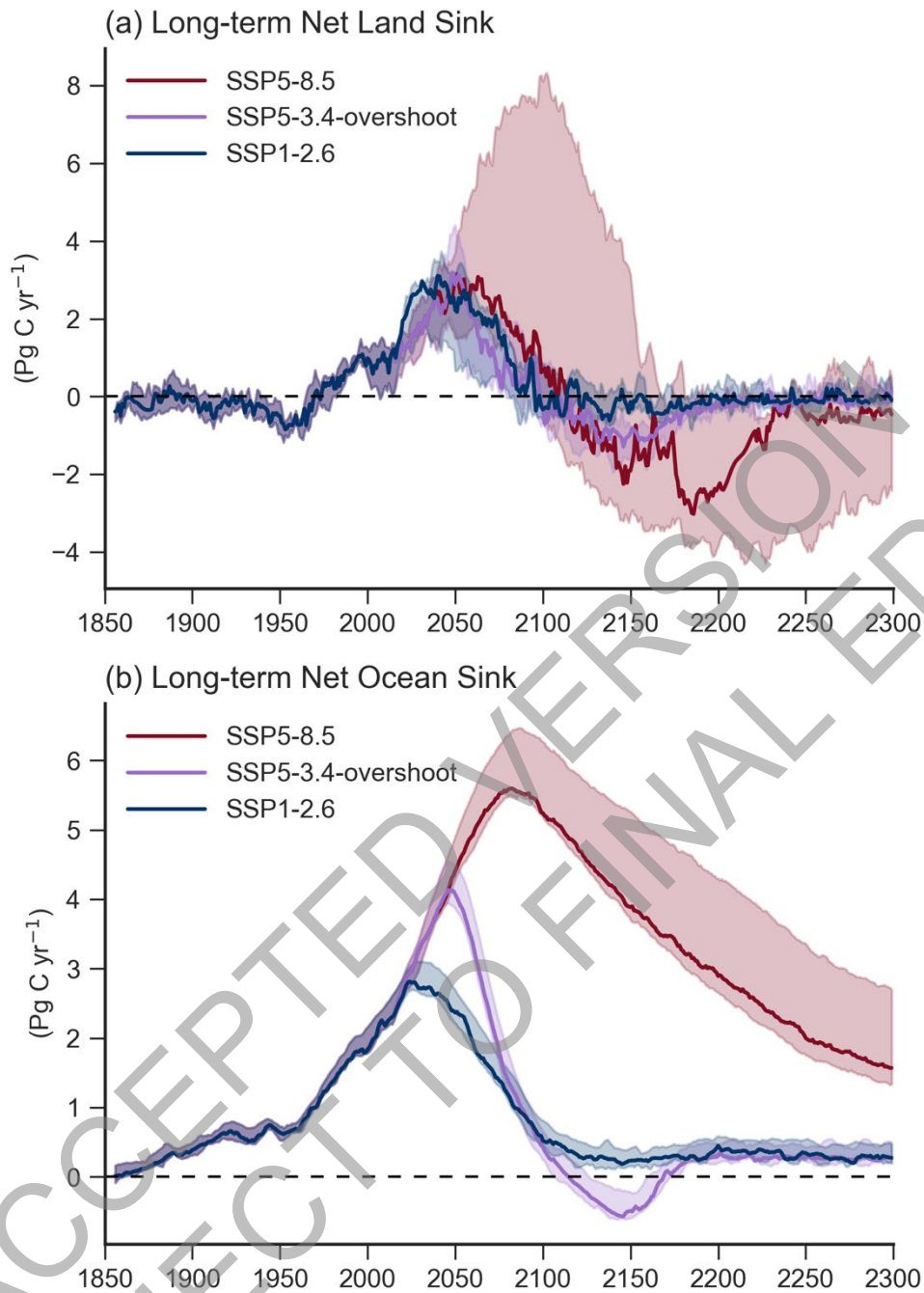
**Figure 5.28: Examples of emergent constraints on the carbon cycle in Earth system models (ESMs), reproduced from previously published studies:** (a) projected global mean atmospheric CO<sub>2</sub> concentration by 2060 under the RCP8.5 emissions scenario against the simulated CO<sub>2</sub> in 2010 (Friedlingstein et al., 2014; Hoffman et al., 2014); (b) sensitivity of tropical land carbon to warming ( $\gamma_{LT}$ ) against the sensitivity of the atmospheric CO<sub>2</sub> growth-rate to tropical temperature variability (Cox et al., 2013; Wenzel et al., 2014); (c) sensitivity of extratropical (30°N–90°N) gross primary production to a doubling of atmospheric CO<sub>2</sub> against the sensitivity of the amplitude of the CO<sub>2</sub> seasonal cycle at Kumkahi, Hawaii to global atmospheric CO<sub>2</sub> concentration (Wenzel et al., 2016); (d) change in high-latitude (30°N–90°N) gross primary production versus trend in high-latitude leaf area index or ‘greenness’ (Winkler et al., 2019); (e) sensitivity of the primary production of the Tropical ocean to climate change versus its sensitivity to ENSO-driven temperature variability (Kwiatkowski et al., 2017); (f) global ocean carbon sink in the 2090s versus the current-day carbon sink in the Southern Ocean. In each case, a red-dot represents a single ESM projection, the grey bar represents the emergent relationship between the y-variable and the x-variable, the blue bar represents the observational estimate of the x-axis variable, and the green bar represents the resulting emergent constraint on the y-axis variable. The thicknesses represent  $\pm$  one standard error in each case. Figure after Cox, (2019). Further details on data sources and processing are available in the chapter data table (Table 5.SM.6).

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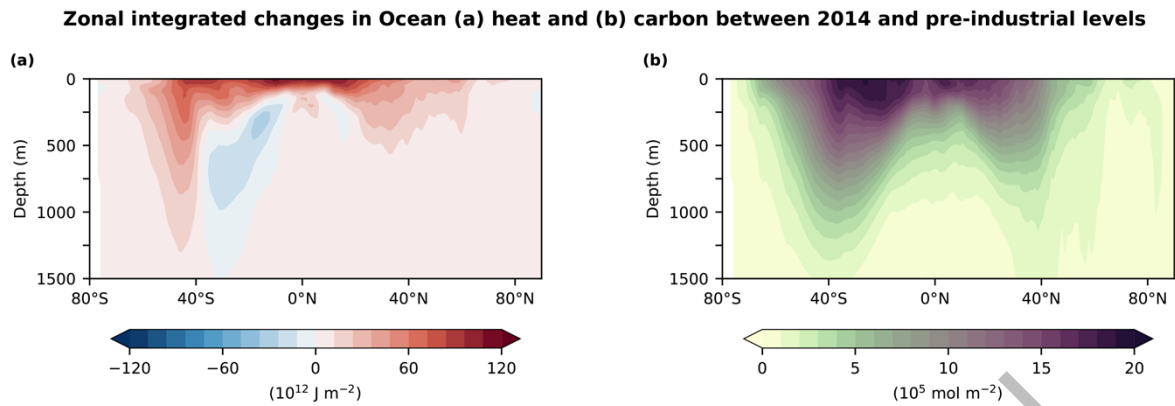
**Figure 5.29: Estimates of the biogeochemical climate feedback parameter ( $\alpha$ ).** The parameter  $\alpha$  (W m<sup>-2</sup> °C<sup>-1</sup>) for a feedback variable  $x$  is defined as  $\alpha_x = \frac{\partial N}{\partial x} \frac{dx}{dT}$  where  $\frac{\partial N}{\partial x}$  is the change in TOA energy balance in response to a change in  $x$  induced by a change in surface temperature ( $T$ ), as in Section 7.4.1.1. (a) Synthesis of biogeochemical feedbacks from panels (b) and (c). Red (blue) bars correspond to positive (negative) feedbacks increasing (decreasing) radiative forcing at the top of the atmosphere. Bars denote the mean and the error bar represents the 5–95<sup>th</sup> percentile range of the estimates; (b) carbon-cycle feedbacks as estimated by coupled carbon-cycle climate models in the CMIP5 (Arora et al., 2013) and CMIP6 (Arora et al., 2020) ensembles, where dots represent single model estimates, and filled (open) circles are those estimates which do (not) include the representation of a terrestrial nitrogen cycle; (c) Estimates of other biogeochemical feedback mechanisms based on various modelling studies. Dots represent single estimates, and coloured bars denote the mean of these estimates with no weighting being made regarding the likelihood of any single estimate, and error bars the 5–95<sup>th</sup> percentile range derived from these estimates. Results in panel (c) have been compiled from (a) Section 5.4.3.2 (Eliseev et al., 2014; Harrison et al., 2018); (b) Section 5.4.3.3 (Schneider von Deimling et al., 2012; Burke et al., 2013; Koven et al., 2015b, 2015a; MacDougall and Knutti, 2016; Burke et al., 2017; Gasser et al., 2018; Kleinen and Brovkin, 2018), where the estimates from Burke et al. have been constrained as assessed in their study (c) Section 5.4.7 (Schneider von Deimling et al., 2012; Koven et al., 2015b; Schneider von Deimling et al., 2015; Turetsky et al., 2020); (d) Section 5.4.7 (Armeth et al., 2010; Denisov et al., 2013; Shindell et al., 2013; Stocker et al., 2013; Zhang et al., 2017); (f) Section 5.4.7 (Xu-Ri et al., 2012; Zaehle, 2013; Stocker et al., 2013; Tian et al., 2019); (g) Section 5.4.7 (Martinez-Rey et al., 2015; Landolfi et al., 2017; Battaglia and Joos, 2018). (h) Section 6.3, Table 6.9 mean and the 5–95<sup>th</sup> percentile range the assessed feedback parameter. Further details on data sources and processing are available in the chapter data table (Table 5.SM.6).

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**Figure 5.30: Trajectories of carbon cycle dynamics for models beyond 2100.** Shown are three scenarios, SSP5–8.5, SSP5–3.4–overshoot, and SSP1–2.6, from four ESMs (CanESM5, UKESM1, CESM2-WACCM, IPSL-CM6a-LR) and one EMIC (UVIC-ESCM, (Mengis et al., 2020)) for which extensions beyond 2100 are available. Solid lines represent the median flux value across the ensemble, and shading represents 15th–85th percentiles across the ensemble. Further details on data sources and processing are available in the chapter data table (Table 5.SM.6).

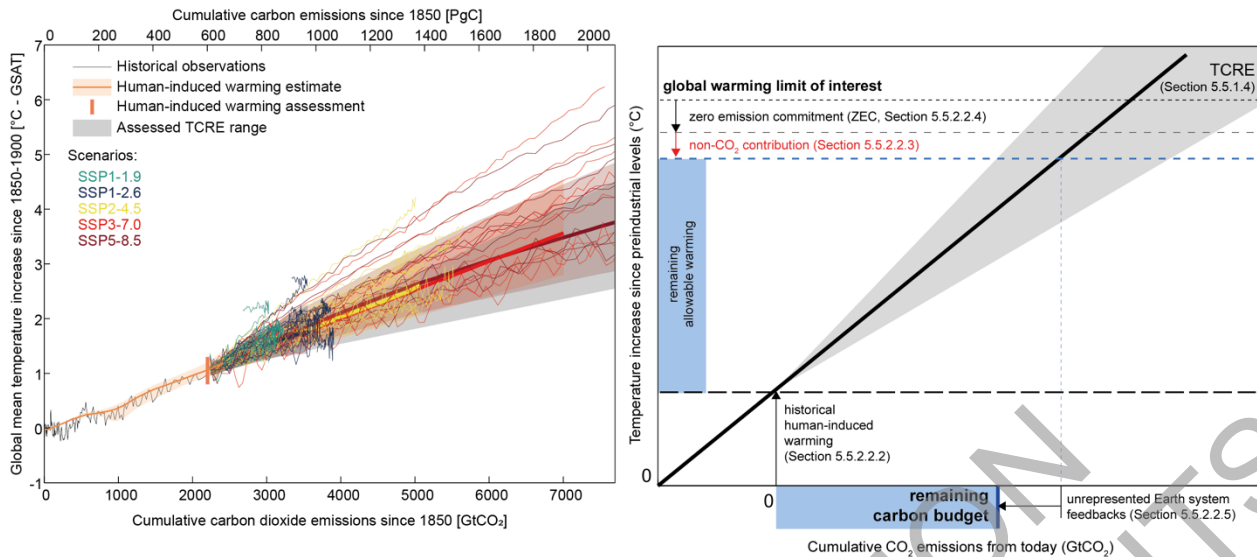
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**Cross-Chapter Box 5.3, Figure 1: CMIP6 multi-model mean of changes in zonally integrated (a) carbon and (b) heat storage in ocean between the pre-industrial and the modern period. Carbon corresponds to dissolved inorganic carbon. Data are shown for the upper 2000m. The modern period is 1995–2014. Adapted from (Frölicher et al., 2015))**

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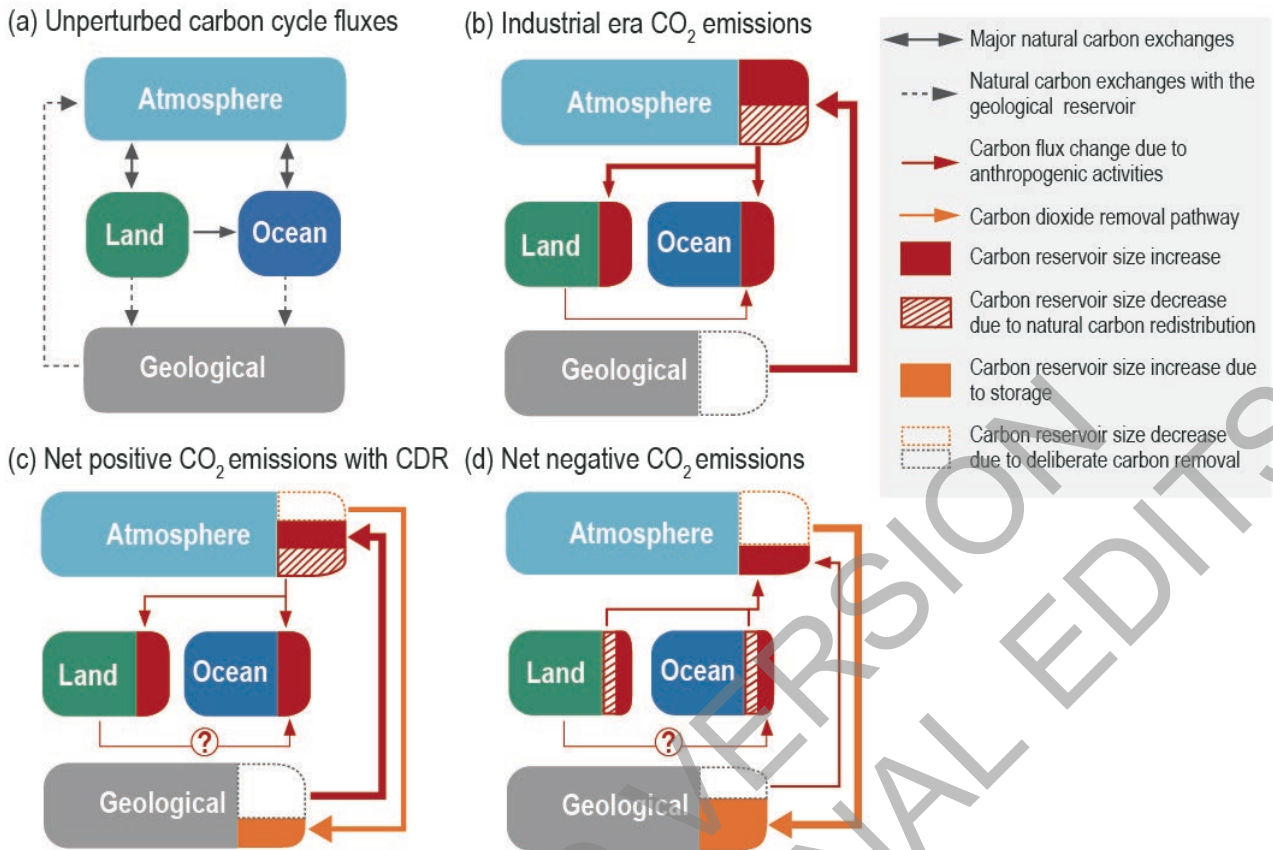




**Figure 5.31: Illustration of relationship between cumulative emissions of carbon dioxide (CO<sub>2</sub>) and global mean surface air temperature increase (left) and conceptual schematic of the assessment of the remaining carbon budget from its constituting components (right).** Carbon budgets consistent with various levels of additional warming are provided in Table 5.8 and should not be read from the illustrations in either panel. Left-hand panel: Historical data (thin black line data) shows historical CO<sub>2</sub> emissions as reported in (Friedlingstein et al., 2020) together with the assessed global mean surface air temperature increase from 1850–1900 as assessed in Chapter 2 (Box 2.3, GSAT). The orange-brown range with its central line shows the estimated human-induced share of historical warming (Haustein et al., 2017). The vertical orange-brown line shows the assessed range of historical human-induced warming for the 2010–2019 period relative to 1850–1900 (Chapter 3). The grey cone shows the assessed range for the transient climate response to cumulative emissions of carbon dioxide (TCRE) assessed to fall *likely* in the 1.0–2.3 °C per 1000 PgC range (Section 5.5.1.4), starting from 2015. Thin coloured lines show CMIP6 simulations for the five scenarios of the AR6 core set (SSP1–1.9, green; SSP1–2.6, blue; SSP2–4.5, yellow; SSP3–7.0, red; SSP5–8.5, maroon), starting from 2015. Diagnosed carbon emissions (Arora et al., 2020) are complemented with estimated land-use change emissions for each respective scenario (Gidden et al., 2018). Coloured areas show the Chapter 4 assessed *likely* range of GSAT projections and thick coloured central lines the median estimate, for each respective scenario, relative to the original scenario emissions (Riahi et al., 2017; Gidden et al., 2018; Rogelj et al., 2018). Right-hand panel: schematic illustration of assessment of remaining carbon budget based on multiple lines of evidence. The remaining allowable warming is estimated by combining the global warming limit of interest with the assessed historical human induced warming (Section 5.5.2.2.2), the assessed future potential non-CO<sub>2</sub> warming contribution (Section 5.5.2.2.3) and the ZEC (Section 5.5.2.2.4). The remaining allowable warming (vertical blue bar) is subsequently combined with the assessed TCRE (Sections 5.5.1.4 and 5.5.2.2.1) and contribution of unrepresented Earth system feedbacks (Section 5.5.2.2.5) to provide an assessed estimate of the remaining carbon budget (horizontal blue bar, Table 5.8). Further details on data sources and processing are available in the chapter data table (Table 5.SM.6).

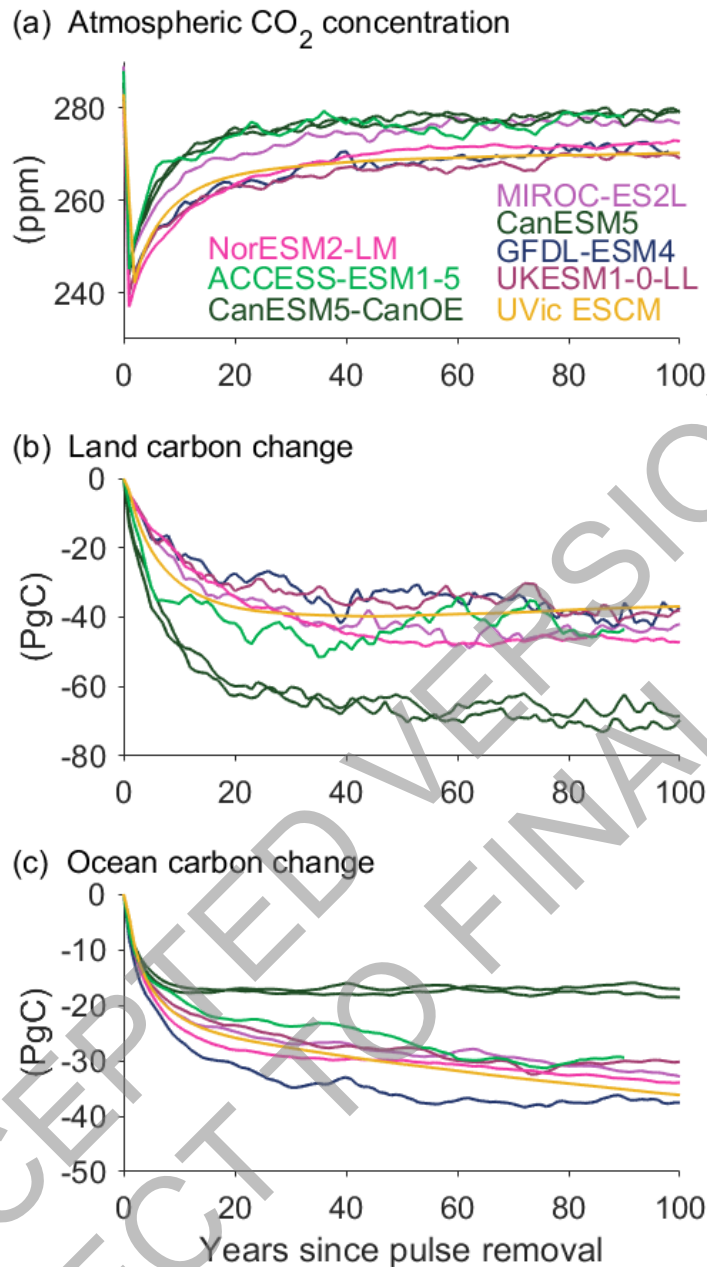
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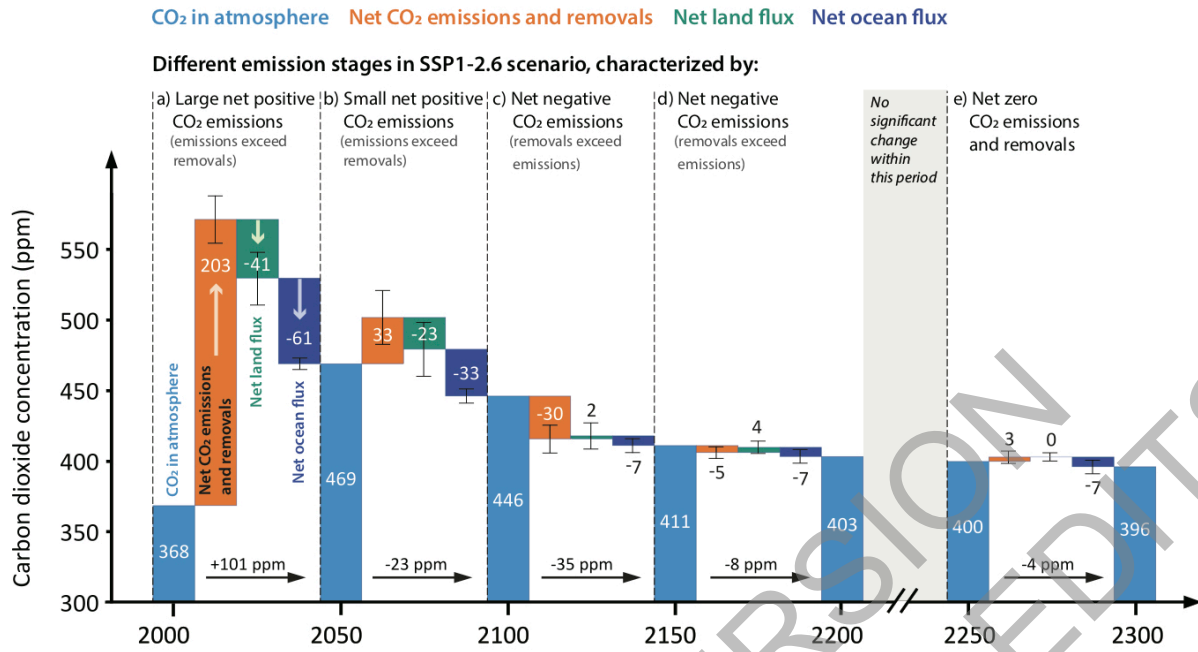
**Box 5.3, Figure 1: Schematic representation of carbon fluxes between atmosphere, land, ocean and geological reservoirs.** Different system conditions are shown: (a) an unperturbed Earth system; and changes in carbon fluxes for (b) an Earth system perturbed by fossil-fuel CO<sub>2</sub> emissions, (c) an Earth system in which fossil-fuel CO<sub>2</sub> emissions are partially offset by CDR, (d) an Earth system in which CDR exceeds CO<sub>2</sub> emissions from fossil fuels (“net negative” CO<sub>2</sub> emissions). Carbon fluxes depicted in (a) (solid and dashed black lines) also occur in (b)-(d). The question mark in the land-to-ocean carbon flux perturbation in (c) and (d) indicates that the effect of CDR on this flux is unknown. Note that box sizes do not scale with the size of carbon reservoirs. Adapted from (Keller et al., 2018a). Further details on data sources and processing are available in the chapter data table (Table 5.SM.6).



**Figure 5.32: Carbon cycle response to instantaneous carbon dioxide (CO<sub>2</sub>) removal from the atmosphere.** (a)

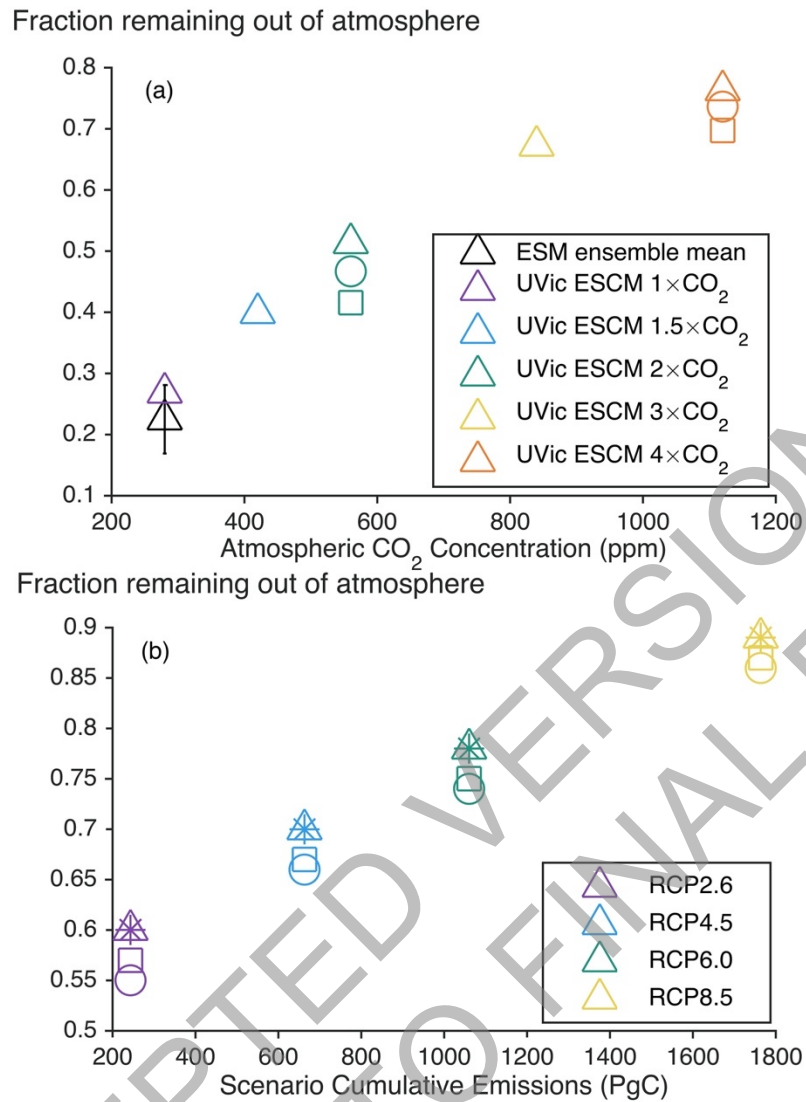
Atmospheric CO<sub>2</sub> concentration, (b) change in land carbon reservoir, (c) change in ocean carbon reservoir. Results are shown for simulations with seven CMIP6 Earth system models and the UVic ESCM model of intermediate complexity forced with 100 PgC instantaneously removed from the atmosphere. The ‘pulse’ removal is applied from a model state in equilibrium with a pre-industrial atmospheric CO<sub>2</sub> concentration (CDRMIP experiment CDR-pi-pulse; Keller et al., 2018b). Changes in land and ocean carbon reservoirs are calculated relative to a pre-industrial control simulation. Data for the UVic ESCM is from Zickfeld et al. (2021). Further details on data sources and processing are available in the chapter data table (Table 5.SM.6).

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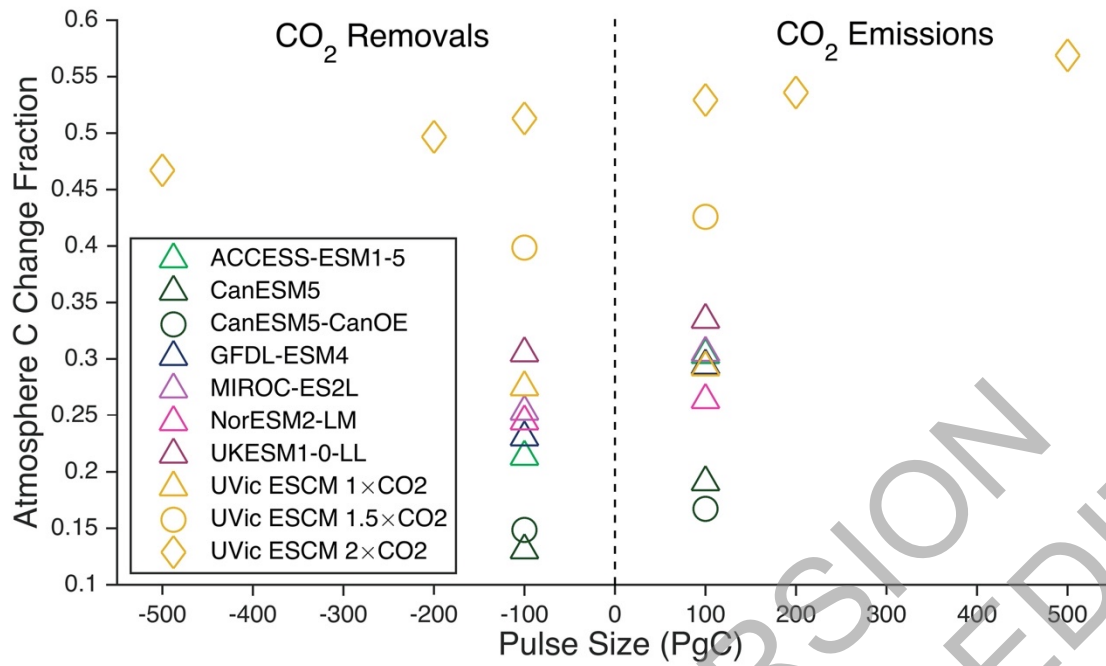
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**Figure 5.33: Carbon sink response in a scenario with net carbon dioxide (CO<sub>2</sub>) removal from the atmosphere.** Shown are CO<sub>2</sub> flux components from concentration-driven Earth system model simulations during different emission stages of SSP1–2.6 and its long-term extension. (a) Large net positive CO<sub>2</sub> emissions, (b) small net positive CO<sub>2</sub> emissions, (c) – (d) net negative CO<sub>2</sub> emissions, (e) net zero CO<sub>2</sub> emissions. Positive flux components act to raise the atmospheric CO<sub>2</sub> concentration, whereas negative components act to lower the CO<sub>2</sub> concentration. Net CO<sub>2</sub> emissions, land and ocean CO<sub>2</sub> fluxes represent the multi-model mean and standard deviation (error bar) of four ESMs (CanESM5, UKESM1, CESM2-WACCM, IPSL-CM6a-LR) and one EMIC (UVic ESCM; (Mengis et al., 2020)). Net CO<sub>2</sub> emissions are calculated from concentration-driven Earth system model simulations as the residual from the rate of increase in atmospheric CO<sub>2</sub> and land and ocean CO<sub>2</sub> fluxes. Fluxes are accumulated over each 50-year period and converted to concentration units (ppm). Further details on data sources and processing are available in the chapter data table (Table 5.SM.6).



**Figure 5.34: Removal effectiveness of carbon dioxide removal (CDR).** (a) Fraction of CO<sub>2</sub> remaining out of the atmosphere for idealised model simulations with CDR applied instantly (pulse removals) from climate states in equilibrium with different atmospheric CO<sub>2</sub> concentration levels (1 to 4 times the pre-industrial atmospheric CO<sub>2</sub> concentration; shown on the horizontal axis). The fraction is calculated 100 years after pulse removal. The black triangle and error bar indicate the multi-model mean and standard deviation for the seven Earth system models shown in Figure 5.32 forced with a 100 PgC pulse removal. Other symbols illustrate results with the UVic ESCM model of intermediate complexity for different magnitudes of pulse removals (triangles: -100 PgC; circles: -500 PgC; squares: -1000 PgC). Data for the UVic ESCM is from (Zickfeld et al., 2021). (b) Perturbation airborne fraction (see text for definition) for model simulations where CDR is applied from four RCPs (shown on the horizontal axis in terms of their cumulative CO<sub>2</sub> emissions during 2020–2099). Symbols indicate results for four CDR scenarios, which differ in terms of the magnitude and rate of CDR (see Jones et al. (2016b) for details). Results are based on simulations with the Hadley Centre Simple Climate–Carbon Model and are shown for the year 2100. Data from Jones et al. (2016b). Further details on data sources and processing are available in the chapter data table (Table 5.SM.6).

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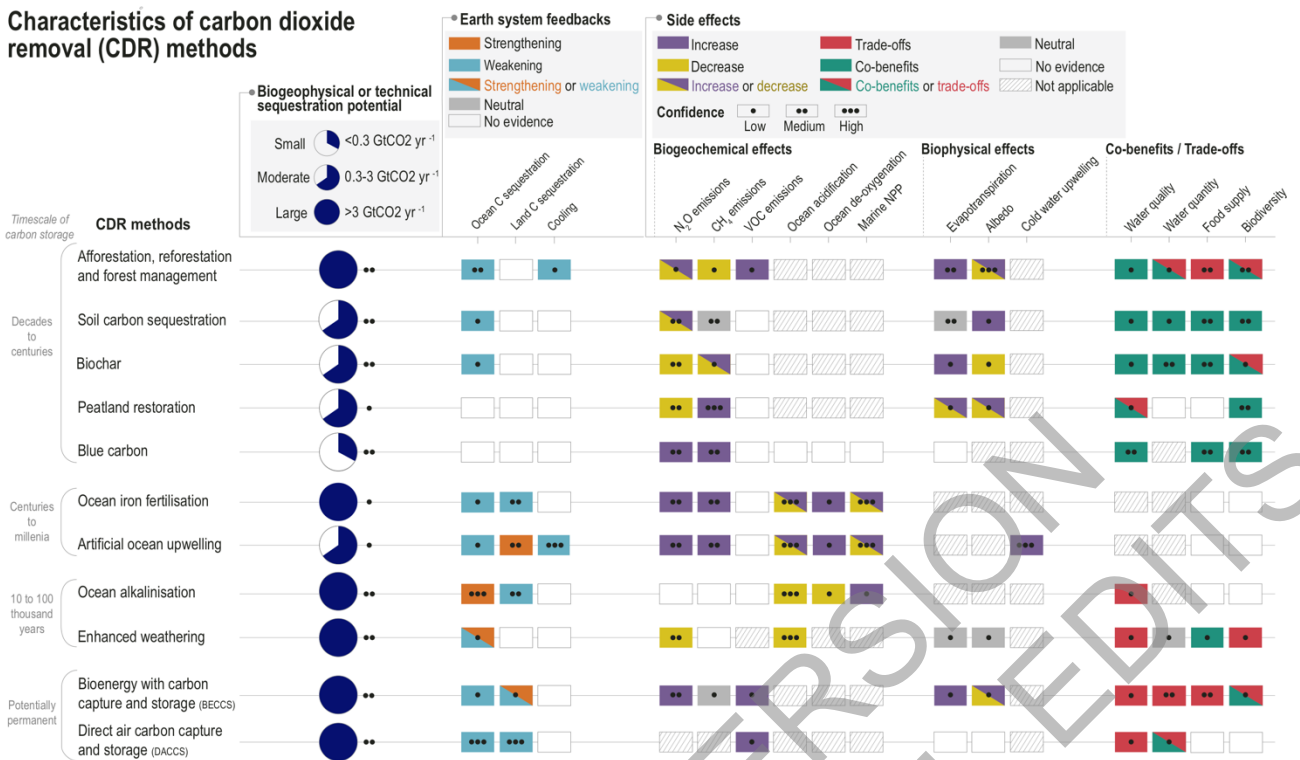


**Figure 5.35: Asymmetry in the atmospheric carbon dioxide (CO<sub>2</sub>) response to CO<sub>2</sub> emissions and removals.**

Shown are the fractions of total CO<sub>2</sub> emissions remaining in the atmosphere (right-hand side) and CO<sub>2</sub> removals remaining out of the atmosphere (left-hand side) 80–100 after a pulse emission/removal. Triangles and green circles denote results for seven Earth system models (ESMs) and the UVic ESCM model of intermediate complexity forced with  $\pm 100$  PgC pulses applied from a pre-industrial state ( $1\times\text{CO}_2$ ) (CDRMIP experiment CDR-pi-pulse; Keller et al. (2018b)). Yellow circles and diamonds indicate UVic ESCM results for CO<sub>2</sub> emissions/removals applied at 1.5 times ( $1.5\times\text{CO}_2$ ) and 2 times ( $2\times\text{CO}_2$ ) the pre-industrial CO<sub>2</sub> concentration, respectively. Pulses applied from a  $2\times\text{CO}_2$  state span the magnitude  $\pm 100$  PgC to  $\pm 500$  PgC. UVic ESCM data is from (Zickfeld et al., 2021). Further details on data sources and processing are available in the chapter data table (Table 5.SM.6).

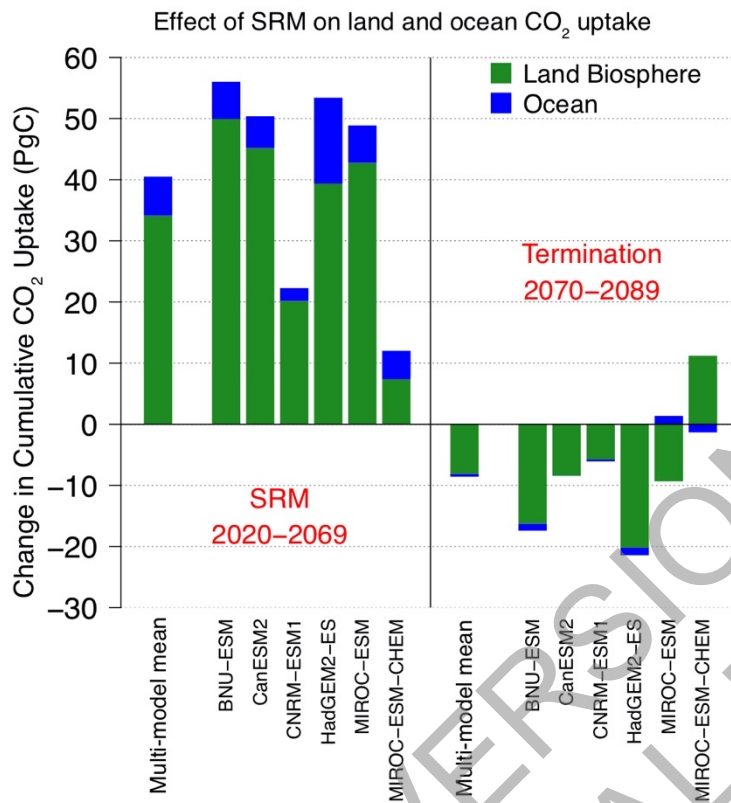
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**Characteristics of carbon dioxide removal (CDR) methods**



**Figure 5.36: Characteristics of carbon dioxide removal (CDR) methods, ordered according to the time scale of carbon storage.** The first column shows biogeophysical (for open-ocean methods) or technical (for all other methods) sequestration potentials (i.e. the sequestration potentials constrained by biological, geophysical, geochemical limits and thermodynamics and, for technical potentials, availability of technologies and practices; technical potentials for some methods also consider social or environmental factors if these represent strong barriers for deployment; see Glossary), classified into low ( $< 0.3 \text{ GtCO}_2 \text{ yr}^{-1}$ ), moderate ( $0.3\text{--}3 \text{ GtCO}_2 \text{ yr}^{-1}$ ) and large ( $>3 \text{ GtCO}_2 \text{ yr}^{-1}$ ) (details underlying this classification are provided in Supplementary Materials Table 5.SM.5). The other columns show Earth system feedbacks that deployment of a given CDR method would have on carbon sequestration and climate, along with biogeochemical, biophysical, and other side effects of a given method. Earth system feedbacks do not include the direct effect of  $\text{CO}_2$  sequestration on atmospheric  $\text{CO}_2$ , only secondary effects. For Earth system feedbacks, the colours indicate whether the feedbacks strengthen or weaken carbon sequestration and the climate cooling effect of a given CDR method. For biogeochemical and biophysical side effects the colours indicate whether the deployment of a CDR method increases or decreases the magnitude of the effect, whereas for co-benefits and trade-offs the colour indicates whether deployment of a CDR method results in beneficial (co-benefits) or adverse side-effects (trade-offs) for water quality and quantity, food production and biodiversity. The details and references underlying the Earth system feedback and side-effect assessment are provided in Supplementary Materials Table 5.SM.4. Further details on data sources and processing are available in the chapter data table (Table 5.SM.6).

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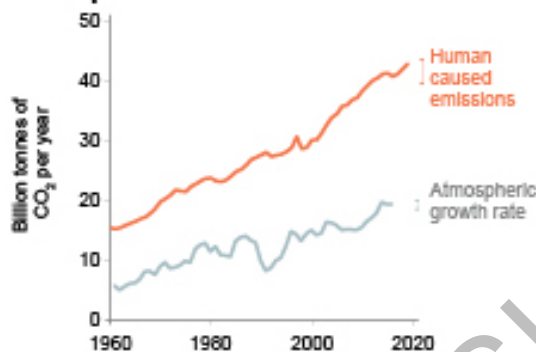
**Figure 5.37: Cumulative carbon dioxide (CO<sub>2</sub>) uptake by land and ocean carbon sinks in response to stratospheric sulfur dioxide (SO<sub>2</sub>) injection.** Results are shown for a scenario with 50-year (2020–2069) continuous stratospheric SO<sub>2</sub> injection at a rate of 5 Tg per year applied to a RCP4.5 baseline scenario (GeoMIP experiment G4; Kravitz et al., (2011)), followed by termination in year 2070. Anomalies are shown relative to RCP4.5 for the multimodel ensemble mean and for each of six ESMs over the 50-year period of stratospheric SO<sub>2</sub> injection (left), and over 20 years after termination of SO<sub>2</sub> injection (right). Adapted from Plazzotta et al. (2019). Further details on data sources and processing are available in the chapter data table (Table 5.SM.6).



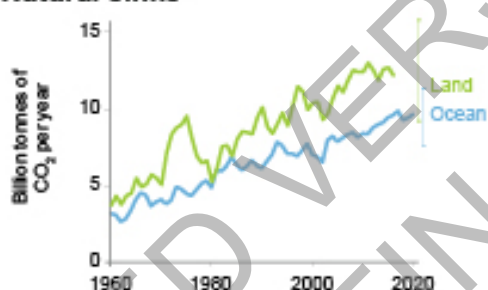
**FAQ 5.1: Is natural removal of carbon from the atmosphere weakening?**

No, natural carbon sinks have taken up a near constant fraction of our carbon dioxide (CO<sub>2</sub>) emissions over the last six decades. However, this fraction is expected to decline in the future if CO<sub>2</sub> emissions continue to increase.

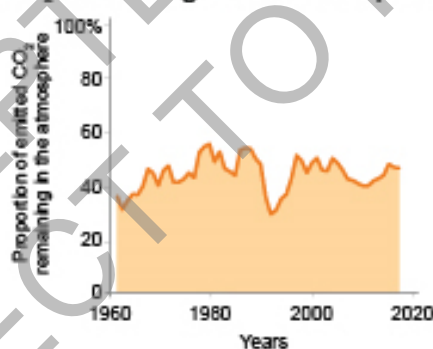
**Atmosphere**



**Natural sinks**



**CO<sub>2</sub> remaining in the atmosphere**



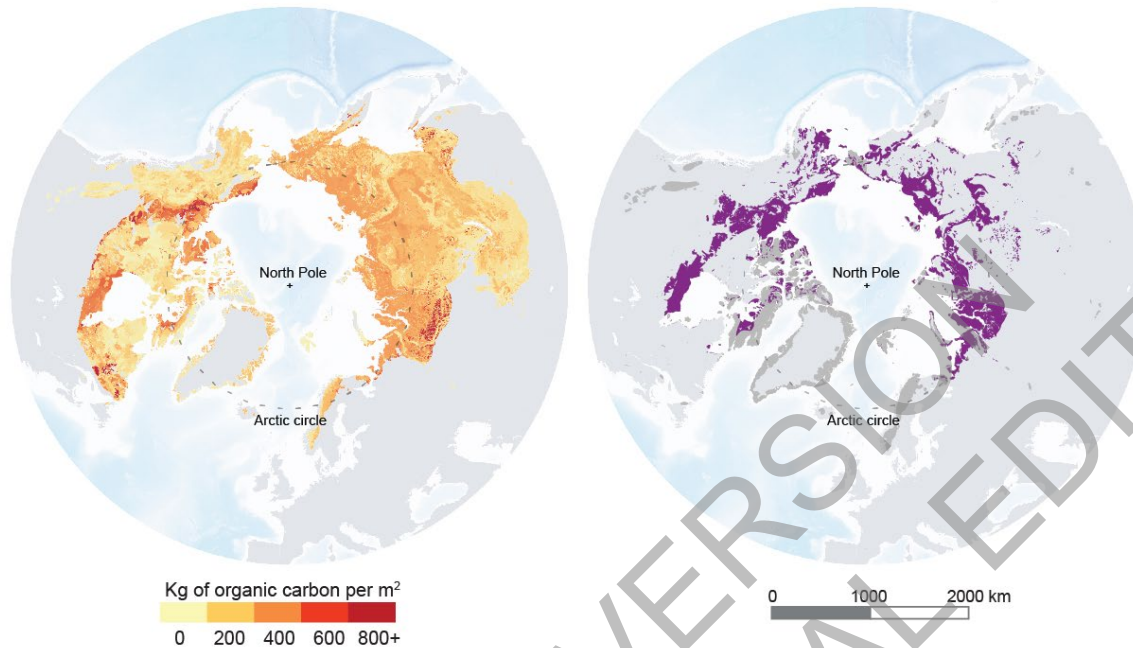
**FAQ 5.1, Figure 1: Atmospheric CO<sub>2</sub> and natural carbon sinks.** (Top) Global emissions of CO<sub>2</sub> from human activities and the growth rate of CO<sub>2</sub> in the atmosphere, (middle) the net land and ocean CO<sub>2</sub> removal (“natural sinks”), as well as (bottom) the fraction of CO<sub>2</sub> emitted by human activities remaining in atmosphere from 1960 to 2019. Lines are the five years running mean, error-bars denote the uncertainty of the mean estimate. See Table 5.SM.6 for more information on the data underlying this figure.

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**FAQ5.2: Can thawing permafrost substantially increase global temperatures?**

The thawing of frozen ground in the Arctic will release carbon that will amplify global warming but this will not lead to runaway warming.

Carbon stored in the Arctic permafrost

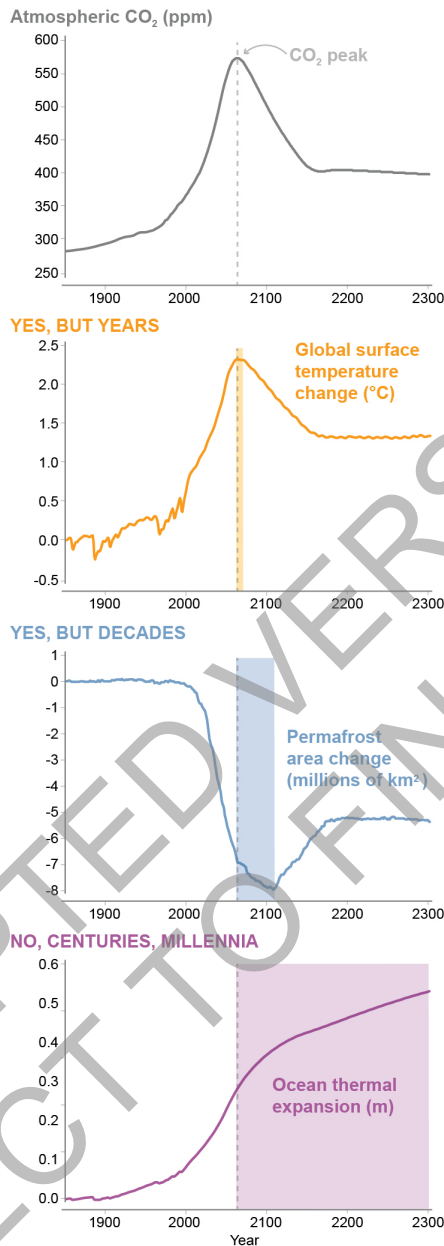
Permafrost **vulnerable** to abrupt thaw

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**FAQ 5.2, Figure 1: The Arctic permafrost is a big pool of carbon that is sensitive to climate change.** (left) Quantity of carbon stored in the permafrost, to 3 m depth (NCSCDV2 dataset) and (right) area of permafrost vulnerable to abrupt thaw (Circumpolar Thermokarst Landscapes dataset).

**FAQ 5.3: Could climate change be reversed by removing CO<sub>2</sub> from the atmosphere?**

Removing more carbon dioxide (CO<sub>2</sub>) from the atmosphere than is emitted into it could reverse some aspects of climate change, but some changes would continue in their current direction for decades to millennia.



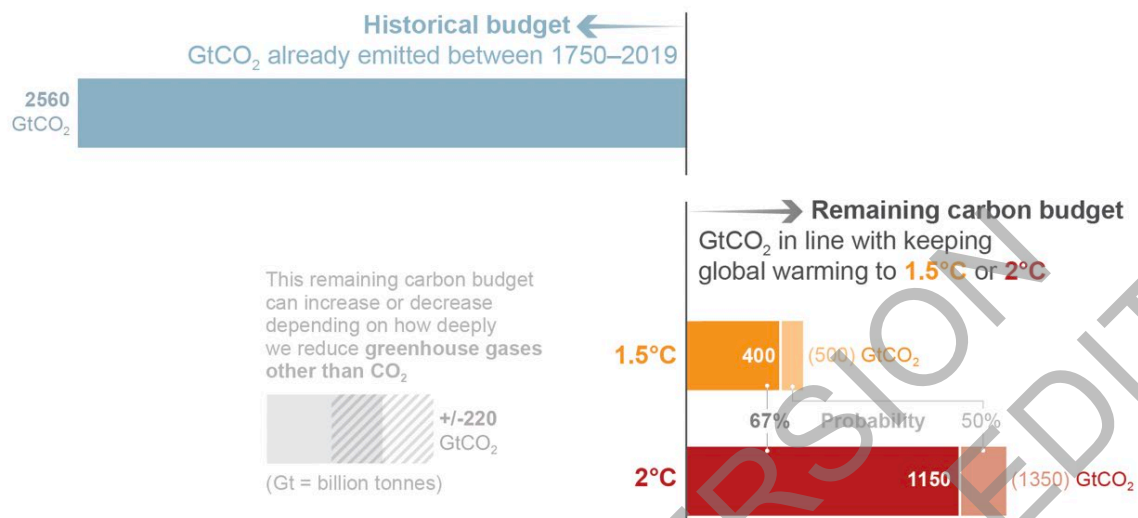
**FAQ 5.3, Figure 1: Changes in aspects of climate change in response to a peak and decline in the atmospheric CO<sub>2</sub> concentration (top panel). The vertical grey dashed line indicates the time of peak CO<sub>2</sub> concentration in all panels. It is shown that the reversal of global surface warming lags the decrease in the atmospheric CO<sub>2</sub> concentration by a few years, the reversal of permafrost area decline lags the decrease in atmospheric CO<sub>2</sub> by decades, and ocean thermal expansion continues for several centuries. Note that the quantitative information in the figure (i.e., numbers on vertical axes) is not to be emphasized as it results from simulations with just one model and will be different for other models. The qualitative behaviour, however, can be expected to be largely model independent.**

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#### FAQ 5.4: What are Carbon Budgets?

The term carbon budget is used in several ways. Most often the term refers to the total net amount of carbon dioxide (CO<sub>2</sub>) that can still be emitted by human activities while limiting global warming to a specified level.

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**FAQ 5.4, Figure 1: Various types of carbon budgets.** Historical cumulative CO<sub>2</sub> emissions determine to a large degree how much the world has warmed to date, while the remaining carbon budget indicates how much CO<sub>2</sub> could still be emitted while keeping warming below specific temperature thresholds. Several factors limit the precision with which the remaining carbon budget can be estimated, and estimates therefore need to specify the probability with which they aim at limiting warming to the intended target level (e.g., limiting warming to 1.5°C with a 67% probability).