

P54/WGI-14 - Changes to the underlying scientific-technical assessment to ensure consistency with the approved SPM
These trickle backs will be implemented in the Chapter during copy-editing

SPM Page:Line	Chapter/Su pp. Material	Chapter Page:Line	Summary of edit to be made
D2.2, 26, 8	6	8:6	to reflect the above change in the text, please add "(high confidence)" between "mitigation" and ","
D2.2, 26, 8	6	8:8	to reflect the above change in the text, please add "(high confidence)" at the end of the sentence "...exceeding the WHO guidelines."
D.2.1, 26, 5	6	75:12	to support the sentence added to SPM.D.2 on non-detection of CO2 growth rate changes, please add "(medium confidence)" to the end of the sentence in chapter 6 CCB, page 75, line 12.
D2.2, 26, 8	6	89:9	to support the high confidence for the new approved sentence in CRP56 - "Scenarios with targeted reductions of air pollutant emissions....GHG emissions with the magnitude of the benefit varying 21 between regions (high confidence).", please add "(high confidence)" to the end of the sentence "...air pollution exceeding the WHO guidelines"

AR6 WGI Report – List of corrigenda to be implemented

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

CHAPTER 6

Document (Chapter, Annex, Supp. Mat...)	Section	Page :Line (based on the final pdf FGD version)	Detailed info on correction to make
6			Replace “climate mitigation” with “climate change mitigation” throughout the chapter
6			Change instances of “VOC” with “NMVOC” throughout the chapter.
6	TOC / 6.6.2.3.4	Numbering of section	Section “6.6.2.3.4” should be “6.6.2.4”
6	TOC / “6.6.2.3.5”	Numbering of section	Section “6.6.2.3.5” should be “6.6.2.5”
6	ES	5:20	Replace “GSAT” by “global surface temperature (GSAT)”
6	ES	6:33	Change “SLCFs” to “SLCF changes”
6	ES	7:43	Replace [0.00–0.10] by [0.00 to 0.10] °C
6	Section 6.1.1	9:28	Replace “account for” with “includes”
6	Section 6.1.1	9:33	Add “net” after or
6	Figure 6.1	9	A new version has been produced with modified arrows in the schematic representation of ozone chemistry modified (uploaded on the figure manager)
6	Section 6.1.1	10:8	Delete “oxidation”
6	Table 6.1	10 (6th row called NMVOCs)	Add a footnote to indicate that some NMVOCs are biogenic volatile organic compounds (BVOCs)
6	6.1.1	11:10	Remove the first “mu” after 50
6	Table 6.1	11 (last row, column 6 called Mineral Dust)	Replace “-” with “+/-”
6	Table 6.1	10-11	Add “Ecosystem” in the 7th Column “Other effects on climate” for SO ₂ , Sulphates, Nitrates, Carbonaceous aerosols, Sea spray and Mineral dust
6	Section 6.1.2	11:39	Delete “ERFaci”
6		12:23	Replace “land-use” by “land use” remove hyphen
6	6.1.2	12:14-23	Check if BVOC and SOA (used in p12 are spelled out before their first use (biogenic volatile organic compounds and secondary organic compounds
6	6.1.3	12:55	“Modelling” should be “models”
6	6.2.1	13:50	Add at the end of the sentence “and in Section 6.7.1.1.”
6	6.2.1	13:55	“timeseries” should be “time serie”
6	6.2.1	13 : 34	Replace “Wang et al., 2014c, 2014b” by “R. Wang et al., 2014; S.X. Wang et al., 2014”
6	6.2.1	13 : 41 14 : 11	Replace “Klimont et al., 2017b” by “Klimont et al., 2017a”
6	6.2.1	14 : 26-27 14 : 29	Replace “Jiang et al., 2018a” by “Jiang et al., 2018”
6	6.2.1	14 : 35	Replace “Liu et al., 2016a” by “F. Liu et al., 2016”
6	6.2.1	14:2	Remove , after i.e.
6	6.2.1	14:4	add , between “emissions” and “making”
6	6.2.1	14:23	Add , before “with”
6	6.2.1	14:36	; should be :
6	6.2.1	14:43	“sector” should be “sectors”

6	Section 6.2.1	14:34	Replace “CEDs.” by “CEDS” (Capital S and no point after)
6	6.2.1	15:47	add a , after “1950”
6	6.2.1	15:50	add a , after “terms”
6	6.2.1	15 : 2 15 : 4	Replace “Li et al., 2019” by “M. Li et al., 2019”
6	6.2.2	16:56	Add “period” after “preindustrial”
6	Section 6.2.1	16:26	Replace “OECD countries” by “countries from the Organisation for Economic Co-operation and Development (OECD)”
6	Section 6.2.2	17:2	Replace “.” after Section 2.2.2 by “ and natural sources of methane and N2O are assessed in Sections 5.2.2 and 5.2.3.”
6	6.2.2.1	17:17	Remove “the” before “CMIP5”
6	6.2.2.1	17:23	Add , before and after “however”
6	6.2.2.2	17:39	Check if first occurrence of ESM
6	6.2.2.2	17:42	ESM should be ESMs
6	6.2.2.3	17:55	Is PFT defined earlier?
6	6.2.2.1	17 : 12	Replace “Allen et al., 2019a” by “D.J. Allen et al., 2019”
6	6.2.2.1	17 : 15	Replace “Lamarque et al., 2013c” by “Lamarque et al., 2013b”
6	6.2.2.3	18 : 34	Replace “Heald and Geddes, 2016a” by “Heald and Geddes, 2016”
6		18:50	Replace “land-use” by “land use” remove hyphen
6	6.2.2.3	18:32	Is LULCC previously defined
6	6.2.2.3	18:48	Add “(“ before low confidence
6		19:7	Replace “land-use” by “land use” remove hyphen
6	6.2.2.4	19:18	Add , after “however”
6	6.2.2.4	19:20	Add , after “manner”
6	6.2.2.5	19:31-34	Check if CCN and POA defined earlier
6	6.2.2.5	19:34	“Wind induced” should be “wind-induced”
6	6.2.2.5	19:51	“Year” should be “yr” (x3 on this line)
6	6.2.2.5	19 : 44	Remove “Burrows et al., 2018”
6	6.2.2.5	20:1	“Predict” should be “project”
6	6.2.2.6	20:9	Sentence should start with “Emissions from”
6	6.2.2.6	20:37	Add , before “consistently” and remove the one after
6	6.2.2.6	20:39	Move possible (line 40) before “leading”
6	6.2.2.6	20:24	Replace “differ” by “by”
6	6.2.2.6	20:54	Replace “in pre-industrial to the 1980s of biomass SLCF emissions” by “SLCF emissions from biomass burning from the preindustrial to 1980s”
6	6.3	21:7	Remove the s to SLCFs
6	6.3	21:8-9	Replace “aqueous” by “heterogeneous”
6	Section 6.2.2.6	21:1	Replace “confidence of” by “confidence in”
6	Box 6.1	22:44	Replace ‘multimodel’ with ‘multi-model’. Add hyphen
6	Box 6.1	22 : 38	Replace “Naik et al., 2013b” by “Naik et al., 2013”
6	6.3.1	23 : 23	Replace “Myhre et al., 2013b” by “Myhre et al., 2013”
6	6.3.1	23 : 29-30 23 : 39	Replace “Zhao et al., 2019b” by “Y. Zhao et al., 2019”
6	Section 6.3.1	23:20	Add “chemical” between “methane” and “lifetime”
6	Section 6.3.1	24:24	Remove the minus sign before “ δ (ln total)/ δ (ln[CH4])”
6	6.3.2.1	25:6	“In troposphere” should be “in the troposphere”
6	6.3.2.1	25 : 21 26 : 13-14	Replace “Myhre et al., 2013b” by “Myhre et al., 2013”
6	Section 6.3.2.1	26:16	Insert space after “109”
6	6.3.2.2	27:46	Are DU and ODS defined somewhere?
6	Section 6.3.2.2	27:38	Delete “-” from “total-column ozone”
6	6.3.3.1	28 : 27	Replace “Lamarque et al., 2013b” by “Lamarque et al., 2013a”
6	6.3.3.1	28 : 38	Replace “Jiang et al., 2018b” by “Jiang et al., 2018”
6	6.3.3.1	28 : 40	Replace “Liu et al., 2016a” by “F. Liu et al., 2016”

6	Figure 6.6 caption	29:8	Replace “Long term” by “Long-term”
6	6.3.3.2	29:36	Replace “emissions datasets” by “emission datasets”
6	Section 6.3.3.2	30:11-12	Remove (“ before “Buchholz”, remove “,” after “.” and add (“ before “2021”. Final text should be “Buchholz et al. (2021)”
6	6.3.3.2	30:5	“Low confidence” should be in italic
6	6.3.3.3	32:17	Replace “HCHO” by “formaldehyde (HCHO)”
6	6.3.3.4	32:52	“upper troposphere and lower stratosphere” by “Upper Troposphere and Lower Stratosphere”
6	Section 6.3.4	34:21	Replace “2015b” with 2015
6	6.3.4	34 : 21	Replace “Leedham Elvidge et al., 2015b” by “Elvidge et al., 2015a”
6	6.3.4	35 : 1-2	Replace “Leedham Elvidge et al., 2015a” by “Elvidge et al., 2015b”
6	6.3.4	35 : 40	Replace “Zhang et al., 2012a” by “F. Zhang et al., 2012”
6	6.3.5	35 : 52 35: 53 35: 55-56 35 : 56	Replace “Gliß et al., 2020” by “Gliß et al., 2021a”
6	6.3.5	35 : 56 36 : 1	Replace “Cherian and Quaas, 2020a” or “Cherian and Quaas, 2020b” by “Cherian and Quaas, 2020”
6	Section 6.3.4	35:2	Replace “2015a” with 2015
6	Figure 6.7	35	A new version has been produced with blurred boundaries between regions (uploaded on the figure manager)
6	6.3.5	35:12	Replace “long term” by “long-term”
6	6.3.5.1	36 : 41	Replace “Myhre et al., 2013b” by “Myhre et al., 2013”
6	6.3.5.1	37 : 47	Replace “Gliß et al., 2020” by “Gliß et al., 2021a”
6	6.3.5.2	38 : 27	Replace “Zhang et al., 2012b” by “H. Zhang et al., 2012”
6	6.3.5.2	38 : 38 38 : 40	Replace “Gliß et al., 2020” by “Gliß et al., 2021b”
6	Section 6.3.5.1	38:1	Replace “(47%)” with “(47 ± 20%)” and “(40%)” with “(40 ± 30%)”
6	Section 6.3.5.1	38:2	Replace “52%” with “52 ± 21%” and “21%” with “21 ± 14%”
6	Section 6.3.5.3	39:19	Replace first instance of “ERF” with “forcing”, replace “ERF of” by “forcing from”
6	6.3.5.3	39 : 19	Replace “Haines et al., 2017a” by “Haines et a., 2017”
6	6.3.5.3	40 : 8	Replace “Lee et al., 2013a” by “Lee et al., 2013”
6	6.3.5.3	40 : 15 40 : 26 40 : 29 40 : 33 40 : 36	Replace “Gliß et al., 2021” by “Gliß et al., 2021b”
6	6.3.5.3	40 : 19	Replace “Morgan et al., 2019” by “Morgan et al., 2020” Replace “Zhao et al., 2019a” by “D. Zhao et al., 2019”
6	6.3.5.3	40 : 23	Replace “Zhang et al., 2017a” by “G. Zhang et al., 2017”
6	6.3.5.3	40 : 30-31	Replace “Lee et al., 2013b” by “Lee et al., 2013” Replace “Wang et al., 2014a” by “Q. Wang et al., 2014”
6	Table 6.7	43	Replace “Zhao et al., 2019b” by “Y. Zhao et al. (2019)”
6	Section 6.3.6	43:11	Delete “stabilized or”
6	Table 6.7	43	Replace “<” with “within” on rows 4 and 15 of the table
6	Section 6.4	44:13	Replace “ERFs” by “effective radiative forcing (ERF)”
6	6.4	44:33-34	Replace “cloud condensation nuclei (CCN) “ by “CCN” (acronym already defined p19)
6	6.4	44 : 30 44 : 39	Replace “Gliß et al., 2021” by “Gliß et al., 2021b”
6	6.4.1	45 : 7	Replace “Myhre et al., 2013b” by “Myhre et al., 2013”

6	6.4.1	45:6	“medium” should be in italic
6	6.4.1	45:4	“confidence” should be in italic
6	Caption Figure 6.10	46 138	Replace “Multi-model mean Effective radiative forcings (ERFs) due to aerosol changes between 1850 and recent-past (1995-2014).” by “Multi-model mean Effective radiative forcings (ERFs) over recent-past (1995-2014) induced by aerosol changes since 1850.”
6	Figure 6.12	47	A new version has been produced with updated estimates of the aerosol contribution (uploaded on the figure manager)
6	6.4.2	47:54	Replace “1.21 (0.90 to 1.51)” by “1.19 (0.81 to 1.58)”
6	Caption Figure 6.12	47	“Contrails and Light absorbing particle on snow and ice (LAP) contributions to ERF and GSAT change are not represented, but their estimates can be seen on Figure 7.6 And 7.7 respectively)”
6	6.4.2	47 : 52	Replace “Myhre et al., 2013b” by “Myhre et al., 2013”
6	6.4.2	48:45	Replace “SLCFs” with “emitted compounds”
6	6.4.2	48:4	Replace “0.45” by “0.44”
6	6.4.2	48:9-10	Replace “-0.29 (-0.57 to 0.0)” by “-0.27 (-0.55 to 0.01)”
6	6.4.2	48:18	Replace “-0.90 (-0.24 to -1.56)” by “-0.94 (-1.63 to -0.25)”
6	6.4.2	48:19	Replace “-0.22” by “-0.23” and “-0.68” by “-0.70”
6	6.4.2	48:23	Replace “0.063 (-0.28 to 0.42)” by “0.11 (-0.20 to 0.42)”
6	6.4.2	48:30	Replace “-0.20 (-0.03 to -0.41)” by “-0.21 (-0.44 to 0.02)”
6	6.4.3	49:29	Insertion: change “...stratospheric between 1979...” to “...stratospheric cooling between 1979...”
6	6.4.3	49:31	Insertion: change “...near-surface air temperature in comparison...” to “...near-surface air temperature responses in comparison...”
6	6.4.3	49:44	Insert comma: change “...understood providing...” to “...understood, providing...”
6	6.4.3	50:32-34	Change “, and resulting climate effects.” to “, and resulting radiative efficacy.”
6	6.4.3	50:1	“high confidence” should be in italic
6	Caption Figure 6.13	50 142	Replace “Multi-model mean surface air temperature response due to aerosol changes between 1850 and recent-past (1995-2014) calculated as the difference between CMIP6 ‘historical’ and AerChemMIP ‘hist-piAer’ experiments, where a) is the spatial pattern of the annual mean surface air temperature response, and b) is the mean zonally averaged response. Model means are derived from years 1995-2014.” By “Multi-model mean surface air temperature response over recent-past (1995-2014) induced by aerosol changes since 1850 calculated as the difference between CMIP6 ‘historical’ and AerChemMIP ‘hist-piAer’ experiments averaged over 1995-2014 years, where a) is the spatial pattern of the annual mean surface air temperature response, and b) is the mean zonally averaged response.”
6	6.4.3	50:4	change “(6.10)” to “(Figure 6.10)”
6	6.4.3	50 : 6	Replace “Sand et al., 2013b” by “Sand et al., 2013a”
6	6.4.3	50 : 25	Replace “Ren et al., 2020b” by “Ren et al., 2020”
6	6.4.3	50 : 28	Replace “Sand et al., 2013a” by “Sand et al., 2013b”
6	6.4.4	51 : 53	Replace “Wang et al., 2018c” by “X. Wang et al., 2018”
6	6.4.5	52 : 49	Replace “Burrows et al., 2018” by “Cochran et al., 2017” https://pubs.acs.org/doi/abs/10.1021/acs.accounts.6b00603
6	6.4.5	53 : 13	Replace “Wang et a., 2018b” by “S. Wang et al., 2018”
6	6.4.5	53:43	change “water vapour and increases” to “water vapour, and increases”

6	6.4.5	54:11	Close parenthesis: "(see Section 6.2.2.3" to "(see Section 6.2.2.3)"
6	6.4.5	54 : 46	Replace "Arneth et al. (2010" by "Arneth et al. (2010a)"
6	6.4.6	55 : 28	Replace "National Academy of Sciences, 2015" by "NRC, 2015"
6	6.4.5, Table 6.8 caption	55:17	Add subscript: Greek alpha should have a subscript "x" like in line 14
6	6.4.6.	56:20	"Marine source brightening by "Marine Source Brightening"
6	6.4.6.	56:40	Cirrus cloud thinning by "Cirrus Cloud Thinning
6	6.4.6.	56:5	Replace "Stratospheric aerosol injections" by "Stratospheric Aerosol Injections (SAI)"
6	6.5.1	57:30	"high confidence" should be in italic
6	6.5.1	57:32	"medium confidence" should be in italic
6	6.4.6	57 : 39	Replace "Banerjee et al., 2015" by "Banerjee et al., 2016"
6	6.5.1	58:50	Replace 'lighting' with 'lightning'. Misspelling.
6	6.5.2	59:25	"no confidence level" should not be in italic
6	Caption Figure 6.14	59:6	"O3", 3 should be in subscript
6	Caption Figure 6.14	59:6	Add "" around ssp370SST
6	Caption Figure 6.14	59:7	Add "" around ssp370pdSST
6	6.5.2	59 : 32	Replace "Wang et al., 2018a; Zhao et al., 2019c" by "B. Wang et al., 2018; Zhao et al., 2019"
6	6.5.2	59 : 37 59 : 46	Replace "Allen et al., 2016b, 2019b" by "R.J. Allen et al., 2016, 2019"
6	6.5.3	60:17-18	"medium confidence" should be in italic
6	6.5.3	60:19	"low confidence" should be in italic
6	6.6	61:16-17	"mitigation of specific anthropogenic sources" should be changed to "mitigation of emissions from specific anthropogenic sources"
6	6.6	61:38-39	Add "to remain below" after "and/or"
6	6.6	61:50-51	"The effects of the measures to contain the spread of COVID-19 in 2020 on air quality and climate are discussed in cross-chapter box 6.1 at the end of this section" should be rephrased as "The air quality and climate effects of the measures to contain the spread of COVID-19 in 2020 are discussed in cross-chapter box 6.1 at the end of this section"
6	6.6	61 : 8	Replace "Shindell et al., 2012a, 2017a" by "Shindell et al., 2012, 2017b"
6	6.6	61 : 26	Replace "Allen et al., 2016a" by "M.R. Allen et al., 2016"
6	6.6	61 : 35-36	Replace "Shindell et al., 2012a, 2017a" by "Shindell et al., 2012, 2017b" Replace "Haines et al. (2017a" by ""Haines et al., 2017)"
6	6.6	61 : 36	Replace "Shindell et al., 2017a" by "Shindell et al., 2017b
6	6.6	61 : 41-42	Replace "Shindell et al., 2017a" by "Shindell et al., 2017b" Replace "Haines et al., 2017a" by "Haines et al., 2017" Replace "Li et al., 2018a" by "Li et al., 2018"
6	6.6	61 : 43-44	Replace "Li et al., 2018a" by "Li et al., 2018"
6	6.6	61 : 45	Replace "Lund et al., 2014a" by "Lund et al., 2014"
6	6.6.1	62:25	Replace "long term" by "long-term"
6	6.6.1	62:26	Replace "long term" by "long-term"
6	6.6.2.1	63:21	in "...(AFOLU) is a significant..." Replace "is" by "are"
6	6.6.2.1	63:36	Change "carbonaceous aerosol" to "carbonaceous aerosols"
6	6.6.2.1	63:24	Replace "near term" by "near-term"
6	6.6.2.2	63:38-39	Replace "The impacts of residential CO and VOC emissions are warming and SO2 and NOx are net cooling." by "The net effect of residential CO and VOC emissions is warming and that of SO2 and NOx is cooling of the atmosphere.
6	6.6.2.2	63:39	Remove "net"
6	6.6.2.2	63:41-44	Replace "Estimates of global residential sector direct aerosol-radiation effects range from -20 to +60 mW m-2 (Kodros et al., 2015) and -66 to +21 mW m-2 (Butt et al., 2016); and aerosol-cloud effects range from -20 to +10 mWm-2 (Kodros et

			al., 2015) and -52 to -16 mW m ⁻² (Butt et al., 2016).” by Estimates of direct aerosol-radiation and aerosol-cloud effects from global residential sector range from -20 to +60 mW m ⁻² (Kodros et al., 2015) and -66 to +21 mW m ⁻² (Butt et al., 2016) and from -20 to +10 mW m ⁻² (Kodros et al., 2015) and -52 to -16 mW m ⁻² (Butt et al., 2016), respectively.”
6	6.6.2.1	63 : 26	Replace “Heald and Geddes, 2016b” by “Heald and Geddes, 2016”
6	6.6.2.2	63 : 52	Replace “Silva et al., 2016a” by “Silva et al., 2016”
6	6.6.2.3.1	64 : 29 64 : 32-33 64: 36 64: 37 64: 39 64: 41 64 : 50	Replace “Lee et al. (2020a)” by “Lee et al. (2021)”
6	6.6.2.2	63:48-49	Delete “The net sign of the impacts of carbonaceous aerosols from residential burning on radiative forcing and climate (warming or cooling) is ambiguous”
6	6.6.2.2	64:1-6	Replace this sentence by “The net climate effect of a one year pulse of current emissions from the residential sector is warming in the near term of +0.0018±0.00084°C from fossil fuel use and +0.0014±0.0012°C from biofuel use. Over a 100 year time horizon, this warming is +0.0017±0.00017°C and +0.0001±0.000079°C, respectively (Lund et al., 2020). This is due to the effects of BC, CH ₄ , CO and VOCs which add to that of CO ₂ but the uncertainty in the sign of carbonaceous aerosol net effects challenges overall quantitative understanding of this sector and leads to low confidence in this assessment.”
6	6.6.2.3.1	64:13	Replace the sentence by “Aviation is associated with a range of SLCFs, in particular emissions of NO _x and aerosol particles, alongside emissions of water vapour and CO ₂ .”
6	6.6.2.2	64:29	Replace “builds on” by “built upon”
6	6.6.2.3.1	64:29	There are three copies of the same Lee et al (2021) paper. 2020a, 2020b and 2021 are the same
6	6.6.2.3.1	64:35-37	Replace the sentence by “Contrails and aviation-induced cirrus yield the largest individual positive ERF followed by CO ₂ and NO _x emissions (Lee et al., 2020a). The confidence level in ERF due to contrails and aviation-induced cirrus is assessed to be low by Chapter 7 (Section 7.3.4.2) due to potential missing processes.”
6	6.6.2.3.1	64:44	Replace “Bickel et al. (2020)” by “(Bickel et al. 2020)”. Remove “which has”
6	6.6.2.3.1	64:45	Replace “confirmed studies indicating” by “confirming”. Replace “Ponater et al. (2005) and Rap et al. (2010).” by “(Ponater et al., 2005; Rap et al., 2010).”
6		64:8	Replace “sulphate aerosol yields” by “sulphate aerosols yield”
6		65:20	High-confidence, remove hyphen
6	6.6.2.3.2	65:42	Replace “one year of” with “a year’s worth of”
6	6.6.2.3.2	65:55	Replace “petrol” with “gasoline”
6	6.6.2.3.2	65 : 36	Replace “Liu et al., 2016b” by “H. Liu et al., 2016”
6	6.6.2.3.4	66:34	Replace “As discussed in” by “As discussed by”
6	6.6.2.3.3	66 : 2-3 66 : 4	Replace “Lund et al., 2014b” by “Lund et al., 2014” Replace “Huang et al., 2020b” by “Y. Huang et al., 2020”
6	6.6.2.3.3	66 : 11	Replace “Silva et al., 2016a” by “Silva et al., 2016”

6	6.6.2.3.3	66 : 13	Replace “Silva et al., 2016b” by “Silva et al., 2016”
6	6.6.2.3.5	68 : 31	Replace “Lee et al. (2020b)” by “Lee et al. (2021)”
6	6.6.3	69 : 14-15	Replace “Shindell et al., 2012a, 2017a” by “Shindell et al., 2012, 2017b” Replace “Haines et al. 2017a” by “Haines et al., 2017”
6	6.6.3	69 : 46	Replace “Lund et al., 2014b” by “Lund et al., 2014”
6	6.6.3	69:35	Replace “LowSLCF” by “lowSLCF”
6	6.6.3.2	70:45	Replace “LowSLCF” by “lowSLCF”
6	6.6.3.1	70 : 33-34	Replace “Myhre et al., 2013b” by “Myhre et al., 2013”
6	6.6.3.1	70 : 35	Replace “Li et al., 2019a, 2020a” by “K. Li et al., 2019, 2020”
6	6.6.3.2	71 : 22	Replace “Klimont et al., 2017c” by “Klimont et al., 2017b”
6	6.6.3.3	71 : 43-44	Replace “Shindell et al., 2012b, 2017a” by “Shindell et al., 2012, 2017b” Replace “Haines et al. 2017a” by “Haines et al., 2017” Replace “Klimont et al. 2017c” by “Klimont et al., 2017b”
6	6.6.3.3	71 : 49	Replace “Haines et al. 2017a” by “Haines et al., 2017”
6	6.6.3.3	72 : 19	Replace “Sand et al., 2013a” by “Sand et al., 2013b”
6	6.6.3.3	72 : 43	Replace “Li et al., 2018b” by “Li et al., 2018”
6	6.6.3.3	73 : 12	Replace “Shindell et al., 2017b” by “Shindell et al., 2017a”
6	6.6.3.3	73 : 28	Replace “Schmale et al., 2014b” by “Schmale et al., 2014a”
6	Box 6.2	73 : 55	Replace “Li et al., 2019a” by “K. Li et al., 2019”
6	6.6.3.3	73:20	Replace “maximum technical mitigation potential for CH4 globally” by “maximum technically feasible reductions (MTFR) for CH4 globally”
6	CCB 6.1	75 : 32-33	Replace “Bauwens et al., 2020b” by “Bauwens et a., 2020” Replace “Venter et al., 2020b” by “Venter et al., 2020”
6	CCB 6.1	75 : 40	Replace “Li et al., 2020b” by “L. Li et al., 2020”
6	CCB 6.1	75 : 45	Replace “Huang et al., 2020a” by “X. Huang et al., 2020”
6		77:2-3	Remove the sentence “Scenarios (...) (Forster et al. 2020)”. The second part of the sentence (which referred to WGIII) has been removed after the 12th of March but the sentence reads like a ghost now.
6	6.7.1.1	78:24	Replace “near term” by “near-term”
6	6.7.1.1	78 : 51	Replace “Pinder et al., 2006” by “Pinder et al., 2007”
6	6.7.1.1	79 : 15	Replace “Klimont et al., 2017b” by “Klimont et al., 2017a”
6	6.7.1.1	80:16	Replace “near term” by “near-term”
6	Section 6.7.1.1	80:7	Replace “high-emission” by “ high-CO2-emission“
6	6.7.1.1	81:13	“21st” st should be in superscript”
6	6.7.1.1	81 : 3	Replace “Haines et al. 2017b” by “Haines et al., 2017” Replace “Klimont et al., 2017c” by “Klimont et al., 2017b”
6	6.7.1.2	82:10	Remove the blank before the full stop.
6	6.7.2 title	83:10	Replace “SLCF emissions” by “changes in SLCF emissions”
6	6.7.2.1 title	83:12	Replace “SLCFs” by “changes in SLCFs”
6	6.7.2.1	85:16	Replace “socio-economic developments” by “socio-economic developments and emission controls induced by policies”
6	6.7.2.1	85:20	Replace “warming of the SLCFs” by “warming induced by changes in SLCFs”
6	6.7.2.1	85:21	“in the mitigation scenarios” by “in the scenario considering strong climate change mitigation”
6	6.7.2.1	85:43	“ of the SLCFs” by “induced by changes in SLCFs”
6	6.7.2.2	85:50	“In the SSP3-7.0 scenario the net effect of SLCFs in all regions is an enhanced warming towards the end of the century. Methane then becomes the dominant SLCF, and Africa is the region contributing the most to predicted global warming in 2100 (0.24°C).” by “In the SSP3-7.0 scenario, the net effect induced by changes in SLCFs in all regions is an enhanced warming towards the end of the century, driven predominantly by change in methane. Africa is the region contributing the most to predicted global warming due to SLCF changes in 2100 (0.24°C).”

6	6.7.2.1	85:18	Replace “SLCFs” by SLCF changes”
6	6.7.2.1	85:21-22	Replace “high-emission” by “ high-CO2-emission“
6	Section 6.7.2.2	85:29	Replace “effective radiative forcing” by “ERF”
6	Section 6.7.3	86:22	Replace “long-lived climate forcers” by LLGHGs
6	Figure 6.7		Figure updated (to add a missing data point that should have been included in the FGD and boundaries blurred). Figure has been uploaded to the figure manager.
6	6.7.3	86:54	Replace “of 21 st century” by “of the 21 st century”
6	6.7.3	86:49	“21 st ” st should be in superscript
6	6.7.3	87:14-15	Replace “near and long term” by “near- and long-term”
6		87:47	Replace “land use” by “land- use” ADD hyphen
6	Caption 6.25	88:20	Replace “LowSLCF” by “lowSLCF”
6	6.7.3	88:51	Replace 0.1 °C by 0.08 °C
6	6.7.3	88:54	Replace 0.1 °C by 0.07 °C
6	6.7.3	88:55	Replace [0.1 to 0.20] by [-0.08 to 0.18]
6	6.7.3	89:1-2	Replace “near and long term” by “near- and long-term”
6	Figure 6.2	129	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.)
6	Figure 6.7	135:1-6	Replace Figure 6.7 with updated figure not indicating country borders.
6	Caption Figure 6.25	155:7	Replace “LowSLCF” by “lowSLCF”
6	Caption Figure 6.26	158:1	Replace “LowSLCF” by “lowSLCF”
6	Caption Figure 6.24	154:4	“Effects of “ should be “effects of changes in ”
6	Caption Figure 6.22	152:3	“Effects of “ should be “effects of changes in ”

Chapter 6: Short-lived climate forcers

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1 **Executive Summary**

2
3 Short-lived climate forcers (SLCFs) affect climate and are, in most cases, also air pollutants. They include
4 aerosols (sulphate, nitrate, ammonium, carbonaceous aerosols, mineral dust and sea spray), which are also
5 called particulate matter (PM), and chemically reactive gases (methane, ozone, some halogenated
6 compounds, nitrogen oxides, carbon monoxide, non-methane volatile organic compounds, sulphur dioxide
7 and ammonia). Except for methane and some halogenated compounds whose lifetimes are about a decade or
8 more, SLCFs abundances are highly spatially heterogeneous since they only persist in the atmosphere from a
9 few hours to a couple of months. SLCFs are either radiatively active or influence the abundances of
10 radiatively active compounds through chemistry (chemical adjustments), and their climate effect occurs
11 predominantly in the first two decades after their emission or formation. They can have either a cooling or
12 warming effect on climate, and they also affect precipitation and other climate variables. Methane and some
13 halogenated compounds are included in climate treaties, unlike the other SLCFs that are nevertheless
14 indirectly affected by climate change mitigation since many of them are often co-emitted with CO₂ in
15 combustion processes. This chapter assesses the changes, in the past and in a selection of possible futures of
16 the emissions and abundances of individual SLCFs primarily on a global and continental scale, and how
17 these changes affect the Earth's energy balance through radiative forcing and feedback in the climate system.
18 The attribution of climate and air-quality changes to emission sectors and regions, and the effects of SLCF
19 mitigations defined for various environmental purposes, are also assessed.
20

21 **Recent Evolution in SLCF Emissions and Abundances**

22
23 **Over the last decade (2010–2019), strong shifts in the geographical distribution of emissions have led**
24 **to changes in atmospheric abundances of highly variable SLCFs (*high confidence*). Evidence from**
25 **satellite and surface observations show strong regional variations in trends of ozone (O₃), aerosols and**
26 **their precursors (*high confidence*). In particular, tropospheric columns of nitrogen dioxide (NO₂) and**
27 **sulphur dioxide (SO₂) continued to decline over North America and Europe (*high confidence*) and to increase**
28 **over South Asia (*medium confidence*), but have declined over East Asia (*high confidence*). Global carbon**
29 **monoxide (CO) abundance has continued to decline (*high confidence*). The concentrations of**
30 **hydrofluorocarbons (HFCs) are increasing (*high confidence*). Global carbonaceous aerosol budgets and**
31 **trends remain poorly characterized due to limited observations, but sites representative of background**
32 **conditions have reported multi-year declines in black carbon (BC) over several regions of the Northern**
33 **Hemisphere. {6.2, 6.3, 2.2.4, 2.2.5, 2.2.6}**
34

35 **There is no significant trend in the global mean tropospheric concentration of hydroxyl (OH) radical –**
36 **the main sink for many SLCFs, including methane (CH₄) – from 1850 up to around 1980 (*low***
37 ***confidence*) but OH has remained stable or exhibited a positive trend since the 1980s (*medium***
38 ***confidence*). Global OH cannot be measured directly and is inferred from Earth system and climate**
39 **chemistry models (ESMs, CCMs) constrained by emissions and from observationally constrained inversion**
40 **methods. There is conflicting information from these methods for the 1980–2014 period. ESMs and CCMs**
41 **concur on a positive trend since 1980 (about a 9% increase over 1980–2014) and there is *medium confidence***
42 **that this trend is mainly driven by increases in global anthropogenic nitrogen oxide (NO_x) emissions and**
43 **decreases in anthropogenic CO emissions. The observation-constrained methods suggest either positive**
44 **trends or the absence of trends based on *limited evidence and medium agreement*. Future changes in global**
45 **OH, in response to SLCF emissions and climate change, will depend on the interplay between multiple**
46 **offsetting drivers of OH. {6.3.6 and Cross-Chapter Box 5.1}**
47

48 **Effect of SLCFs on Climate and Biogeochemical Cycles**

49 **Over the historical period, changes in aerosols and their ERF have primarily contributed to a surface**
50 **cooling, partly masking the greenhouse gas driven warming (*high confidence*). Radiative forcings**
51 **induced by aerosol changes lead to both local and remote temperature responses (*high confidence*). The**
52 **temperature response preserves the South-North gradient of the aerosol ERF – hemispherical asymmetry- but**
53 **is more uniform with latitude and is strongly amplified towards the Arctic (*medium confidence*). {6.4.1,**
54 **6.4.3}**

55 **Since the mid-1970s, trends in aerosols and their precursor emissions have led to a shift from an**

1 **increase to a decrease of the magnitude of the negative globally-averaged net aerosol ERF (*high***
2 ***confidence*)**. However, the timing of this shift varies by continental-scale region and has not occurred for
3 some finer regional scales. The spatial and temporal distribution of the net aerosol ERF from 1850 to 2014 is
4 highly heterogeneous, with stronger magnitudes in the Northern Hemisphere (*high confidence*). {6.4.1}

5
6 **For forcers with short lifetimes (e.g., months) and not considering chemical adjustments, the response**
7 **in surface temperature occurs strongly as soon as a sustained change in emissions is implemented, and**
8 **that response continues to grow for a few years, primarily due to thermal inertia in the climate system**
9 **(*high confidence*)**. Near its maximum, the response slows down but will then take centuries to reach
10 equilibrium (*high confidence*). For SLCFs with longer lifetimes (e.g., a decade), a delay equivalent to their
11 lifetimes is appended to the delay due to thermal inertia (*high confidence*). {6.6.1}

12
13 **Over the 1750-2019 period, changes in SLCF emissions, especially of CH₄, NO_x and SO₂, have**
14 **substantial effects on effective radiative forcing (ERF) (*high confidence*)**. The net global emissions-based
15 ERF of NO_x is negative and that of non-methane volatile organic compounds (NMVOCs) is positive, in
16 agreement with the AR5 assessment (*high confidence*). For methane, the emission-based ERF is twice as
17 high as the abundance-based ERF (*high confidence*) attributed to chemical adjustment mainly via ozone
18 production. SO₂ emission changes make the dominant contribution to the ERF from aerosol–cloud
19 interactions (*high confidence*). Over the 1750-2019 period, the contributions from the emitted compounds to
20 GSAT changes broadly match their contributions to the ERF (*high confidence*). Since a peak in emissions-
21 induced SO₂ ERF has already occurred recently and since there is a delay in the full GSAT response,
22 changes in SO₂ emissions have a slightly larger contribution to GSAT change than for CO₂ emissions,
23 relative to their respective contributions to ERF. {6.4.2, 6.6.1 and 7.3.5}

24
25 **Reactive nitrogen, ozone and aerosols affect terrestrial vegetation and the carbon cycle through**
26 **deposition and effects on large scale radiation (*high confidence*)**. However, the magnitude of these effects
27 on the land carbon sink, ecosystem productivity and hence their indirect CO₂ forcing remain uncertain due to
28 the difficulty in disentangling the complex interactions between the individual effects. As such, these effects
29 are assessed to be of second order in comparison to the direct CO₂ forcing (*high confidence*), but effects of
30 ozone on terrestrial vegetation could add a substantial (positive) forcing compared with the direct ozone
31 forcing (*low confidence*). {6.4.5}

32
33 **Climate feedbacks induced from changes in emissions, abundances or lifetimes of SLCFs mediated by**
34 **natural processes or atmospheric chemistry are assessed to have an overall cooling effect (*low***
35 ***confidence*)**, that is, a total negative feedback parameter, of **-0.20 [-0.41 to +0.01] W m⁻² °C⁻¹**. These
36 non-CO₂ biogeochemical feedbacks are estimated from ESMs, which have advanced since AR5 to include a
37 consistent representation of biogeochemical cycles and atmospheric chemistry. However, process-level
38 understanding of many chemical and biogeochemical feedbacks involving SLCFs, particularly natural
39 emissions, is still emerging, resulting in *low confidence* in the magnitude and sign of most of SLCF climate
40 feedback parameter. {6.2.2, 6.4.5}

41 42 **Future Projections for Air Quality Considering Shared Socio-economic Pathways (SSPs)**

43
44 **Future air quality (in term of surface ozone and PM concentrations) on global to local scales will be**
45 **primarily driven by changes in precursor emissions as opposed to climate change (*high confidence*)**
46 **and climate change is projected to have mixed effects**. A warmer climate is expected to reduce surface O₃
47 in regions remote from pollution sources (*high confidence*) but is expected to increase it by a few parts per
48 billion over polluted regions, depending on ozone precursor levels (*medium to high confidence*). Future
49 climate change is expected to have mixed effects, positive or negative, with an overall low effect, on global
50 surface PM and more generally on the aerosol global burden (*medium confidence*), but stronger effects are
51 not excluded in regions prone to specific meteorological conditions (*low confidence*). Overall, there is *low*
52 *confidence* in the response of surface ozone and PM to future climate change due to the uncertainty in the
53 response of the natural processes (e.g., stratosphere–troposphere exchange, natural precursor emissions,
54 particularly including biogenic volatile organic compounds, wildfire-emitted precursors, land and marine
55 aerosols, and lightning NO_x) to climate change. {6.3, 6.5}

1 **The SSPs span a wider range of SLCF emissions than the Representative Concentration Pathways,**
2 **representing better the diversity of future options in air pollution management (*high confidence*).** In
3 the SSPs, the socio-economic assumptions and climate mitigation ambition primarily drive future emissions,
4 but the SLCF emission trajectories are also steered by varying levels of air pollution control originating from
5 the SSP narratives, independently from climate mitigation. Consequently, SSPs consider a large variety of
6 regional ambition and effectiveness in implementing air pollution legislation and result in wider range of
7 future air pollution levels and SLCF-induced climate effects. {6.7.1}

8
9 **Air pollution projections range from strong reductions in global surface ozone and PM (e.g., SSP1-2.6,**
10 **with strong mitigation of both air pollution and climate change) to no improvement and even**
11 **degradation (e.g., SSP3-7.0 without climate change mitigation and with only weak air pollution**
12 **control) (*high confidence*).** Under the SSP3-7.0 scenario, PM levels are projected to increase until 2050
13 over large parts of Asia, and surface ozone pollution is projected to worsen over all continental areas through
14 2100 (*high confidence*). Without climate change mitigation but with stringent air pollution control (SSP5-
15 8.5), PM levels decline through 2100, but high methane levels hamper the decline in global surface ozone at
16 least until 2080 (*high confidence*). {6.7.1}

17 **Future Projections of the Effect of SLCFs on GSAT in the Core SSPs**

18
19
20 **In the next two decades, it is *very likely* that the SLCF emission changes in the WGI core set of SSPs**
21 **will cause a warming relative to 2019, whatever the SSPs, in addition to the warming from long-lived**
22 **greenhouse gases. The net effect of SLCF and HFC changes on GSAT across the SSPs is a *likely***
23 **warming of 0.06°C–0.35°C in 2040 relative to 2019. Warming over the next two decades is quite**
24 **similar across the SSPs due to competing effects of warming (methane, ozone) and cooling (aerosols)**
25 **SLCFs.** For the scenarios with the most stringent climate and air pollution mitigations (SSP1-1.9 and SSP1-
26 2.6), the *likely* near-term warming from the SLCFs is predominantly due to sulphate aerosol reduction, but
27 this effect levels off after 2040. In the absence of climate change policies and with weak air pollution control
28 (SSP3-7.0), the *likely* near-term warming due to changes in SLCFs is predominantly due to increases in
29 methane, ozone and HFCs, with smaller contributions from changes in aerosols. SSP5-8.5 has the highest
30 SLCF-induced warming rates due to warming from methane and ozone increases and reduced aerosols due
31 to stronger air pollution control compared to the SSP3-7.0 scenario. {6.7.3}

32
33 **At the end of the century, the large diversity of GSAT response to SLCFs among the scenarios**
34 **robustly covers the possible futures, as the scenarios are internally consistent and span a range from**
35 **very high to very low emissions.** In the scenarios without climate change mitigation (SSP3-7.0, SSP5-8.5),
36 the *likely* range of the estimated warming due to SLCFs in 2100 relative to 2019 is 0.4°C–0.9°C {6.7.3,
37 6.7.4}. In SSP3-7.0 there is a near-linear warming due to SLCFs of 0.08°C per decade, while for SSP5-8.5
38 there is a more rapid warming in the first half of the century. For the scenarios considering the most stringent
39 climate and air pollution mitigations (SSP1-1.9 and SSP1-2.6), the reduced warming from reductions in
40 methane, ozone and HFCs partly balances the warming from reduced aerosols, and the overall SLCF effect is
41 a *likely* increase in GSAT of 0.0°C–0.3°C in 2100, relative to 2019. The SSP2-4.5 scenario (with moderate
42 climate and air pollution mitigations) results in a *likely* warming in 2100 due to the SLCFs of 0.2°C–0.5°C,
43 with the largest warming from reductions in aerosols. {6.7.3}

44 **Potential Effects of SLCF Mitigation**

45
46
47 **Over time scales of 10 to 20 years, the global temperature response to a year's worth of current**
48 **emissions of SLCFs is at least as large as that due to a year's worth of CO₂ emissions (*high confidence*).**
49 **Sectors producing the largest SLCF-induced warming are those dominated by CH₄ emissions: fossil**
50 **fuel production and distribution, agriculture and waste management (*high confidence*).** On these time
51 scales, SLCFs with cooling effects can significantly mask the CO₂ warming in the case of fossil fuel
52 combustion for energy and land transportation, or completely offset the CO₂ warming and lead to an overall
53 net cooling in the case of industry and maritime shipping (prior to the implementation of the revised fuel-
54 sulphur limit policy for shipping in 2020) (*medium confidence*). Ten years after a one-year pulse of present-
55 day aviation emissions, SLCFs induce strong, but short-lived warming contributions to the GSAT response

1 (*medium confidence*), while CO₂ both gives a warming effect in the near term and dominates the long-term
2 warming impact (*high-confidence*). {6.6.1, 6.6.2}

3
4 **The effects of the SLCFs decay rapidly over the first few decades after pulse emission. Consequently,**
5 **on time scales longer than about 30 years, the net long-term global temperature effects of sectors and**
6 **regions are dominated by CO₂ (*high confidence*).** The global mean temperature response following a
7 climate mitigation measure that affects emissions of both short- and long-lived climate forcers depends on
8 their atmospheric decay times, how fast and for how long the emissions are reduced, and the inertia in the
9 climate system. For the SLCFs including methane, the rate of emissions drives the long-term global
10 temperature effect, as opposed to CO₂ for which the long-term global temperature effect is controlled by the
11 cumulative emissions. About 30 years or more after a one-year emission pulse occurs, the sectors
12 contributing the most to global warming are industry, fossil fuel combustion for energy and land
13 transportation, essentially through CO₂ (*high confidence*). Current emissions of SLCFs, CO₂ and N₂O from
14 East Asia and North America are the largest regional contributors to additional net future warming on both
15 short- (*medium confidence*) and long-time scales (*high confidence*). {6.6.1, 6.6.2}

16
17 **At present, emissions from the residential and commercial sectors (fossil and biofuel use for cooking**
18 **and heating) and the energy sector (fossil fuel production, distribution and combustion) contribute the**
19 **most to the world population's exposure to anthropogenic fine PM (*high confidence*), whereas**
20 **emissions from the energy and land transportation sectors contribute the most to ozone exposure**
21 **(*medium to high confidence*).** The contribution of different emission sectors to PM varies across regions,
22 with the residential sector being the most important in South Asia and Africa, agricultural emissions
23 dominating in Europe and North America, and industry and energy production dominating in Central and
24 East Asia, Latin America and Middle East. Energy and industry are important PM_{2.5} contributors in most
25 regions, except Africa (*high confidence*). Source contributions to surface ozone concentrations are similar for
26 all regions. {6.6.2}

27
28 **Assuming implementation and efficient enforcement of both the Kigali Amendment to the Montreal**
29 **Protocol on Ozone Depleting Substances and current national plans limit emissions (as in SSP1-2.6),**
30 **the effects of HFCs on GSAT, relative to 2019, would remain below +0.02°C from 2050 onwards**
31 **versus about +0.04–0.08°C in 2050 and +0.1–0.3°C in 2100 considering only national HFC regulations**
32 **decided prior to the Kigali Amendment (as in SSP5-8.5) (*medium confidence*).** Further improvements in
33 the efficiency of refrigeration and air-conditioning equipment during the transition to low-global-warming-
34 potential refrigerants would bring additional GHG reductions (*medium confidence*) resulting in benefits for
35 climate change mitigation and to a lesser extent for air quality due to reduced air pollutant emissions from
36 power plants. {6.6.3, 6.7.3}

37
38 **Future changes in SLCFs are expected to cause an additional warming. This warming is stable after**
39 **2040 in scenarios leading to lower global air pollution as long as methane emissions are also mitigated,**
40 **but the overall warming induced by SLCF changes is higher in scenarios in which air quality**
41 **continues to deteriorate (induced by growing fossil fuel use and limited air pollution control) (*high***
42 ***confidence*).** If a strong air pollution control resulting in reductions in anthropogenic aerosols and non-
43 methane ozone precursors was considered in SSP3-7.0, it would lead to a *likely* additional near-term global
44 warming of 0.08 [0.00–0.10] °C in 2040. An additional concomitant methane mitigation (consistent with
45 SSP1's stringent climate mitigation policy implemented in the SSP3 world) would not only alleviate this
46 warming but would turn this into a cooling of 0.07 with a *likely* range of [-0.02 to 0.14] °C (compared with
47 SSP3-7.0 in 2040). Across the SSPs, the collective reduction of CH₄, ozone precursors and HFCs can make a
48 difference of GSAT of 0.2 with a *very likely* range of [0.1–0.4] °C in 2040 and 0.8 with a *very likely* range
49 of [0.5–1.3] °C at the end of the 21st century (comparing SSP3-7.0 and SSP1-1.9), which is substantial in the
50 context of the Paris Agreement. Sustained methane mitigation, wherever it occurs, stands out as an option
51 that combines near- and long-term gains on surface temperature (*high confidence*) and leads to air quality
52 benefits by reducing surface ozone levels globally (*high confidence*). {6.6.3, 6.7.3, 4.4.4}

53
54 **Rapid decarbonization strategies lead to air quality improvements but are not sufficient to achieve, in**
55 **the near term, air quality guidelines set for fine PM by the World Health Organization (WHO),**

1 **especially in parts of Asia and in some other highly polluted regions (*high confidence*)**. Additional CH₄
2 and BC mitigation would contribute to offsetting the additional warming associated with SO₂ reductions that
3 would accompany decarbonization (*high confidence*). Strong air pollution control as well as strong climate
4 change mitigation, implemented separately, lead to large reductions in exposure to air pollution by the end of
5 the century (*high confidence*). Implementation of air pollution controls, relying on the deployment of
6 existing technologies, leads more rapidly to air quality benefits than climate change mitigation, which
7 requires systemic changes. However, in both cases, significant parts of the population are projected to remain
8 exposed to air pollution exceeding the WHO guidelines. Additional policies envisaged to attain Sustainable
9 Development Goals (SDG) (e.g., access to clean energy, waste management) bring complementary SLCF
10 reduction. Only strategies integrating climate, air quality, and developments goals are found to effectively
11 achieve multiple benefits. {6.6.3, 6.7.3, Box 6.2}

12 **Implications of COVID-19 Restrictions for Emissions, Air Quality and Climate**

13 **Emissions reductions associated with COVID-19 containment led to a discernible temporary**
14 **improvement of air quality in most regions, but changes to global and regional climate are**
15 **undetectable above internal variability**. Global anthropogenic NO_x emissions decreased by a maximum of
16 about by 35% in April 2020 (*medium confidence*). There is *high confidence* that, with the exception of
17 surface ozone, these emission reductions have contributed to improved air quality in most regions of the
18 world. Global fossil CO₂ emissions decreased by 7% (with a range of 5.8% to 13.0%) in 2020 relative to
19 2019, largely due to reduced emissions from the transportation sector (*medium confidence*). Overall, the net
20 ERF, relative to ongoing trends, from COVID-19 restrictions was likely small and positive for 2020 (less
21 than 0.2 W m⁻²), thus temporarily adding to the total anthropogenic climate influence, with positive forcing
22 from aerosol changes dominating over negative forcings from CO₂, NO_x and contrail cirrus changes.
23 Consistent with this small net radiative forcing, and against a large component of internal variability, Earth
24 system model simulations show no detectable effect on global or regional surface temperature or
25 precipitation (*high confidence*). {Cross-Chapter Box 6.1}

6.1 Introduction

Short lived climate forcers (SLCFs) are a set of chemically and physically reactive compounds with atmospheric lifetimes typically shorter than two decades but differing in terms of physiochemical properties and environmental effects. SLCFs can be classified as direct or indirect, with direct SLCFs exerting climate effects through their radiative forcing and indirect SLCFs being precursors of direct climate forcers. Direct SLCFs include methane (CH₄), ozone (O₃), short lived halogenated compounds, such as hydrofluorocarbons (HFCs), hydrochlorofluorocarbons (HCFCs), and aerosols. Indirect SLCFs include nitrogen oxides (NO_x), carbon monoxide (CO), non-methane volatile organic compounds (NMVOCs), sulphur dioxide (SO₂), and ammonia (NH₃). Aerosols consist of sulphate (SO₄²⁻), nitrate (NO₃⁻), ammonium (NH₄⁺), carbonaceous aerosols (e.g., black carbon (BC), organic aerosols (OA)), mineral dust, and sea spray (see Table 6.1) and can be present as internal or external mixtures and at sizes from nano-meters to tens of micro-meters. SLCFs can be emitted directly from natural systems and anthropogenic sources (primary) or can be formed by reactions in the atmosphere (secondary) (Figure 6.1).

6.1.1 Importance of SLCFs for climate and air quality

The atmospheric lifetime determines the spatial and temporal variability, with most SLCFs showing high variability, except CH₄ and many HCFCs and HFCs that are also well mixed (as a consequence CH₄ is discussed together with other well-mixed GHGs in Chapters 2, 5, and 7). In contrast to well-mixed greenhouse gases, such as CO₂, CH₄ and some HFCs, the radiative forcing effects of most of the SLCFs are largest at regional scales and climate effects predominantly occur in the first two decades after their emissions or formation. However, changes in their emissions can also induce long-term climate effects, for instance by altering some biogeochemical cycles. Therefore, the temporal evolution of radiative effects of SLCFs follows that of emissions, i.e., when SLCF emissions decline to zero their atmospheric abundance and radiative effects decline towards zero. The total influence of individual SLCF emissions on radiative forcing and climate account for their effects on the abundances of other forcers through chemistry (chemical adjustments).

SLCFs can affect climate by interacting with radiation or by perturbing other components of the climate system (e.g., the cryosphere and carbon cycle through deposition, or the water cycle through modifications of cloud properties via cloud condensation nuclei or ice nuclei). SLCFs can have either net warming or cooling effects on climate. In addition to altering the Earth's radiative balance, many SLCFs are also air pollutants with adverse effects on human health and ecosystems. SLCFs are of interest for climate policies (e.g., CH₄, HFCs), and are regulated as air pollutants (e.g., aerosols, O₃) or because of their deleterious influence on stratospheric ozone (e.g., HCFCs). The list of SLCFs assessed in this chapter and their effects are provided in Table 6.1.

[START FIGURE 6.1 HERE]

Figure 6.1: Sources and processes leading to atmospheric short-lived climate forcer (SLCF) burden and their interactions with the climate system. Both direct and indirect SLCFs and the role of atmospheric processes for the lifetime of SLCFs are depicted. Anthropogenic emission sectors illustrated are fossil fuel exploration, distribution and use, biofuel production and use, waste, transport, industry, agricultural sources, and open biomass burning. Emissions from natural systems include those from open biomass burning, vegetation, soil, oceans, lightning, and volcanoes. SLCFs interact with solar or terrestrial radiation, surface albedo, and cloud or precipitation system. The radiative forcing due to individual SLCF can be either positive or negative. Climate change induces changes in emissions from most natural systems as well as from some anthropogenic emission sectors (e.g. agriculture) leading to a climate feedback (purple arrows). Climate change also influences atmospheric chemistry processes, such as chemical reaction rates or via circulation changes, thus affecting atmospheric composition leading to a climate feedback. Air pollutants influence emissions from terrestrial vegetation, including agriculture (the grey arrow).

1 [END FIGURE 6.1 HERE]

2
3
4 [START TABLE 6.1 HERE]

5
6 **Table 6.1:** Overview of SLCFs of interest for Chapter 6. For each SLCF, its source types, lifetime in the atmosphere, and associated radiatively active agent is given. Source type can be primary (emitted from source categories) and/or secondary (formed through atmospheric oxidation mechanisms in the atmosphere). Unless otherwise noted, the stated lifetime refers to tropospheric lifetime*. Climate effect of increased SLCFs is indicated as “+” for warming and “-“ for cooling. “Direct” is used for SLCFs exerting climate effects through their radiative forcing and “Indirect” for SLCFs which are precursors affecting the atmospheric burden of other climatically active compounds. Other processes through which SLCFs affect climate are listed where applicable. The World Health Organization (WHO) guidelines for air quality (AQ) are given, where applicable, to show which SLCFs are regulated for air quality purposes.

Compounds	Source Type ^a	Lifetime	Direct	Indirect	Climate Forcing	Other effects on climate	WHO AQ guidelines ^b
CH ₄	Primary	~9 years ~12 years (perturbation time)	CH ₄	O ₃ , H ₂ O, CO ₂	+		No ^c
O ₃	Secondary	Hours - weeks	O ₃	CH ₄ , secondary organic aerosols, sulphates	+	Ecosystem	100 µg m ⁻³ 8-hour mean
NO _x (= NO + NO ₂)	Primary	Hours - days		O ₃ , nitrates, CH ₄	+/-	Ecosystem	40 µg m ⁻³ annual mean 200 µg m ⁻³ 1-hour mean
CO	Primary + Secondary	1-4 months		O ₃ , CH ₄	+		No
NMVOCs	Primary + Secondary	Hours - months		O ₃ , CH ₄ , organic aerosols	+/-		No
SO ₂	Primary	Days (trop.) to weeks (strat.)		sulphates, nitrates, O ₃	-		20 µg m ⁻³ 24-hour mean 500 µg m ⁻³ 10-minute mean
NH ₃	Primary	Hours		Ammonium Sulphate, Ammonium Nitrate	-	Ecosystem	No
HCFCs	Primary	Months – years	HCFCs	Stratospheric O ₃	+/-		No ^c
HFCs	Primary	Days – years	HFCs		+		No ^c
Halons and Methylbromide	Primary	Years	Halons and Methylbromide	Stratospheric O ₃	+/-		No ^c
Very Short-Lived halogenated Species (VSLs)	Primary	less than 0.5 year		Stratospheric O ₃	-		No ^c
Sulphates	Secondary	Minutes – weeks	Sulphates		-	Cloud	as part of PM ^d
Nitrates	Secondary	Minutes – weeks	Nitrates		-	Cloud	as part of PM ^d
Carbonaceous aerosols	Primary + Secondary	Minutes to Weeks	BC, OA		+/-	Cryo, Cloud	as part of PM ^d

Sea spray	Primary	day to week	Sea spray		-	Cloud	as part of PM ^d
Mineral dust	Primary	Minutes to Weeks	Mineral dust		-	Cryo, Cloud	as part of PM ^d

* For lifetimes reported in this table, it is assumed that the compounds are uniformly mixed throughout the troposphere, however, this assumption is unlikely for compounds with lifetimes < 1 year and therefore, the reported values should be viewed as approximations (Prather et al., 2001).

^a Source types can be primary (emitted from source categories) and/or secondary (formed by reactions in the atmosphere). Cryo: effect on planetary albedo through deposition on snow and ice; ^b Krzyzanowski and Cohen (2008); ^c regulated through Kyoto/Montreal protocol; ^d For Particulate Matter with diameter < 2.5 µm (PM_{2.5}): 10 µg m⁻³ annual mean or 25 µg m⁻³ 24 hour mean (99th percentile) and for Particulate Matter with diameter < 10 µm (PM₁₀): 20 µg m⁻³ annual mean or 50 µg m⁻³ 24 hour mean (99th percentile)

[END TABLE 6.1 HERE]

As depicted in Figure 6.1, emissions of SLCFs are governed by anthropogenic activities and sources from natural systems (see Section 6.2 for details). Atmospheric chemistry in this context is both a source and a sink of SLCFs. For instance, O₃ and secondary aerosols are exclusively formed through atmospheric mechanisms (Section 6.3.2 and 6.3.5 respectively). The hydroxyl (OH) radical, the most important oxidizing agent in the troposphere, acts as a sink for SLCFs by reacting with them, and thereby, influencing their lifetime (Section 6.3.6). Through SLCF radiative forcing and feedbacks (see Section 6.4), key climate parameters, such as temperature, hydrological cycle, and weather patterns are perturbed. Climate change also influences air quality (Section 6.5). As depicted in Figure 6.1, SLCFs affect both climate and air quality, hence SLCF mitigation has linkages to both the issues (Section 6.6). Socio-economic narratives including air quality policies determine future projections of SLCFs in the five core Socioeconomic Pathways (SSPs): SSP1-1.9, SSP1-2.6, SSP2-4.5, SSP3-7.0, and SSP5-8.5 (described in Chapter 1), and in addition, a subset of SSP3 scenarios allows to isolate the effect of various SLCF mitigation trajectories on climate and air quality (Section 6.7).

6.1.2 Treatment of SLCFs in previous assessments

Although O₃, aerosols and their precursors have been considered in previous IPCC assessment reports, AR5 considered SLCFs as a specific category of climate relevant compounds but referred to them as near-term climate forcers (NTCFs) (Myhre et al., 2013). In AR5, the linkages between air quality and climate change were also considered in a more detailed and quantitative way than in previous reports (Kirtman et al., 2013; Myhre et al., 2013).

AR5 WGI assessed radiative forcings for short-lived gases, aerosols, aerosol precursors and aerosol cloud interactions (ERF_{aci}) as well as the evolution of confidence level in the forcing mechanisms from SAR to AR5. Whereas the forcing mechanisms for ozone and aerosol-radiation interactions were estimated to be characterised with *high confidence*, the ones induced by aerosols through other processes remained of *very low to low confidence*. AR5 also reported that forcing agents such as aerosols and ozone changes are highly heterogeneous spatially and temporally and these patterns affect global and regional temperature responses as well as other aspects of climate response such as the hydrologic cycle (Myhre et al., 2013).

AR5 WGI also evaluated the air quality-climate interaction through the projected trends of surface O₃ and PM_{2.5}. Kirtman et al. (2013) concluded with *high confidence* that the response of air quality to climate-driven changes is more uncertain than the response to emission-driven changes, and also that locally higher surface temperatures in polluted regions will trigger regional feedbacks in chemistry and local emissions that will increase peak levels of O₃ and PM_{2.5} (*medium confidence*).

In the Special Report on Global Warming of 1.5 °C (SR15)(Allen et al., 2018a), Rogelj et al., (2018a) state

1 that the evolution of CH₄ and SO₂ emissions strongly influences the chances of limiting warming to 1.5°C,
2 and that, considering mitigation scenarios to limit warming to 1.5 or 2°C, a weakening of aerosol cooling
3 would add to future warming in the near term, but can be tempered by reductions in methane emissions (*high*
4 *confidence*). In addition, as some SLCFs are co-emitted alongside CO₂, especially in the energy and transport
5 sectors, low CO₂ scenarios, relying on decline of fossil fuel use, can result in strong abatement of some
6 cooling and warming SLCFs (Rogelj et al., 2018a).

7
8 On the other hand, specific reductions of the warming SLCFs (CH₄ and BC) would, in the short term,
9 contribute significantly to the efforts of limiting warming to 1.5°C. Reductions of BC and CH₄ would have
10 substantial co-benefits improving air quality and therefore limit effects on human health and agricultural
11 yields. This would, in turn, enhance the institutional and socio-cultural feasibility of such actions in line with
12 the United Nations' Sustainable Development Goals (Coninck et al., 2018).

13
14 Following SR15, the IPCC Special Report Climate Change and Land (SRCCL)(IPCC, 2019a) took into
15 consideration the emissions on land of three major SLCFs: mineral dust, carbonaceous aerosols (BC and
16 OA) and BVOCs (Jia et al., 2019). The SRCCL concluded that: i) there is no agreement about the direction
17 of future changes in mineral dust emissions; ii) fossil fuel and biomass burning, and SOA from natural
18 BVOC emissions are the main global sources of carbonaceous aerosols whose emissions are expected to
19 increase in the near future due to possible increases in open biomass burning and increase in SOA from
20 oxidation of BVOCs (*medium confidence*); iii) BVOCs are emitted in large amounts by forests and they are
21 rapidly oxidised in the atmosphere to form less volatile compounds that can condense and form SOA, and in
22 a warming planet, BVOC emissions are expected to increase but magnitude is unknown and will depend on
23 future land use change, in addition to climate (*limited evidence, medium agreement*).

24
25 Finally, the IPCC Special Report on the Ocean and Cryosphere in a Changing Climate (SROCC)(IPCC,
26 2019b) discussed the effects of BC deposition on snow and glaciers, concluding that there is *high confidence*
27 that darkening of snow through the deposition of BC and other light absorbing particles enhances snow melt
28 in the Arctic (Meredith et al., 2019), but that there is *limited evidence* and *low agreement* that long-term
29 changes in glacier mass of high mountain areas are linked to light absorbing particles (Hock et al., 2019).

30 31 32 **6.1.3 Chapter Roadmap**

33
34 Figure 6.2 presents the Chapter 6 roadmap.

35
36 Specific aspects of SLCFs can also be found in other chapters of this report: the evolution of ozone, HFCs,
37 aerosols as well as long term evolution of methane, dust, volcanic aerosols are discussed in Chapter 2, near-
38 term climate projections and SLCFs are discussed in Chapter 4, the global budget of methane is addressed in
39 Chapter 5, aerosol-cloud and aerosol-precipitation interactions are treated in Chapters 7 and 8, respectively,
40 the global radiative forcing of SLCFs is assessed in Chapter 7, some aspects of downscaling methodology in
41 climate modelling concerning SLCFs are discussed in Chapter 10. The WGII report assesses how climate
42 change affects air pollution and impacts on human health and the WGIII report assesses the role of SLCFs in
43 abatement strategies and their cost effectiveness, the implications of mitigation efforts on air pollution as
44 well as the articulation between air pollution policies and GHG mitigation.

45
46 This chapter discusses air quality from a global point of view with focus on surface ozone and particulate
47 matter surface concentrations. Local and indoor air pollution, as well as effect of air pollution on health, are
48 beyond the scope of this chapter. This assessment is mainly based on results and studies relying on global
49 models or observation datasets operated through global networks or from satellites. Global chemistry-climate
50 models allow to quantify the changes in background concentrations such as surface ozone due to large scale
51 changes in climate or CH₄ by considering comprehensively the physiochemical processes (see Box 6.1). In
52 addition, climate effects are often non-linear responses to concentrations which already respond non-linearly
53 to emissions, with per mass unit effects often larger in pristine than in polluted regions, justifying the
54 relevance of global models. However, specific aspects of urban air quality cannot be captured by global
55 models and require high resolution modelling which reproduce the temporal and spatial variability of

1 emissions and abundances necessary to precisely account for the non-linearity of the chemistry and the
2 sensitivity of local air pollution to its drivers. Consequently, the sectoral analysis in Section 6.6 and the
3 mitigation effects in Section 6.7 cannot be directly applied for local air quality planning.
4

5 Due to their short lifetimes, SLCF trends and effects are strongly related to the localisation and evolution of
6 the emission sources. To better link the drivers of emission evolution and SLCFs, Chapter 6 makes use of
7 regions defined by the WGIII in most of the analysis. An exception is made for the effect of SLCFs on the
8 climate, for which analysis relies on WGI Atlas regions.
9

10
11 **[START FIGURE 6.2 HERE]**

12
13 **Figure 6.2:** Chapter roadmap. See Section 6.1.3 for additional description of the chapter.
14

15 **[END FIGURE 6.2 HERE]**
16

17 18 **6.2 Global and regional temporal evolution of SLCF emissions**

19
20 SLCF emissions originate from a variety of sources driven by anthropogenic activities and natural processes.
21 The natural sources include vegetation, soil, fire, lightning, volcanoes, and oceans. Changes in SLCF
22 emissions from natural systems occur either due to human activities, such as land-use change, or due to
23 global changes. Their sensitivity to climate change thus induces a climate feedback (see Section 6.4.6 for a
24 quantification of these feedbacks). This section reviews the current understanding of historical emissions for
25 anthropogenic, natural, and open biomass burning sources. A detailed discussion of methane sources, sinks,
26 trends and their evaluation are provided in Chapter 5; Section 5.2.2.
27

28 29 **6.2.1 Anthropogenic sources**

30
31 Estimates of global anthropogenic SLCF emissions and their historical evolution that were used in AR5
32 (CMIP5) (Lamarque et al., 2010) have been revised for use in CMIP6 (Hoesly et al., 2018). The update
33 considered new data and assessment of the impact of the environmental policies, primarily regarding air
34 pollution control (Wang et al., 2014c, 2014b; Montzka et al., 2015; Crippa et al., 2016; Turnock et al., 2016;
35 Klimont et al., 2017a; Zanutta et al., 2017; Prinn et al., 2018). Additionally, Hoesly et al. (2018) have
36 extended estimates of anthropogenic emissions back to 1750 and developed an updated and new set of
37 spatial proxies allowing for more differentiated (source sector-wise) gridding of emissions (Feng et al.,
38 2020). The CMIP6 emission inventory has been developed with the Community Emissions Data System
39 (CEDS) that improves upon existing inventories with a more consistent and reproducible methodology,
40 similar to approaches used in, for example, EDGAR (Crippa et al., 2016) and GAINS model (Amann et al.,
41 2011; Klimont et al., 2017b; Höglund-Isaksson et al., 2020) where emissions of all compounds are
42 consistently estimated using the same emission drivers and propagating individual components (activity data
43 and emission factors) separately to capture fuel and technology trends affecting emission trajectories over
44 time. This contrasts the approach used to establish historical emissions for CMIP5 where different data sets
45 available at the time were combined. The CMIP6 exercise is based on the first release of CEDS emission
46 dataset (version 2017-05-18, sometimes referred to hereafter as CMIP6 emissions) whose main features
47 regarding SLCFs are described hereafter. The CEDS has been and will be regularly updated and extended;
48 the recent update of CEDS (Hoesly et al., 2019) and consequences for this assessment is discussed when
49 necessary. Some details on how SLCF emissions have been represented in scenarios used by IPCC
50 assessments can be found in Chapter 1 (see Chapter 1, Section 1.6.1 and Cross Chapter Box 1.4).
51

52 For most of the SLCF species, the global and regional anthropogenic emission trends developed for CMIP6
53 for the period 1850 to 2000 are not substantially different from those used in CMIP5 (Figures 6.18 & 6.19)
54 despite the different method used to derive them. Hoesly et al. (2018, CEDS) developed an independent time
55 series capturing trends in fuel use, technology, and level of control, whereas CMIP5 combined different

1 emission data sets. However, for the period after 1990, the CMIP6 dataset shows for all species, except for
2 SO₂ and CO and since 2011 for NO_x, a different trend than CMIP5, i.e., continued strong growth of
3 emissions driven primarily by developments in Asia (Figure 6.19). The unprecedented growth of East and
4 South Asian emissions since 2000 changed the global landscape of emissions making Asia the dominant
5 SLCF source region (Figures 6.3, 6.19). The Representative Concentration Pathways (RCP) scenarios used
6 in AR5 started from the year 2000 (van Vuuren et al., 2011) and did not represent well the development of
7 SLCF emissions until 2015. The CEDS inventory (Hoesly et al., 2018) includes improved representation of
8 these trends and the estimate for 2014. These findings have been largely supported by several independent
9 emission inventory studies and remote sensing data analysis, however, for the last decade the decline of
10 Asian emissions of SO₂ and NO_x appears underestimated while growth of BC and OC emissions in Asia and
11 Africa seems overestimated in CMIP6, compared to most recent regional evaluations (Klimont et al., 2017b;
12 Zheng et al., 2018b; e.g., Elguindi et al., 2020; Kanaya et al., 2020; McDuffie et al., 2020), which are largely
13 considered in the updated release of the CEDS (Hoesly et al., 2019). Consequently, global CMIP6
14 anthropogenic emissions for 2015 are likely overestimated by about 10% for SO₂ and NO_x and about 15%
15 for BC and OC.

16
17 For SO₂, independent emission inventories and observational evidence show that on a global scale strong
18 growth of Asian emissions has been countered by reduction in North America and Europe (Reis et al., 2012;
19 Amann et al., 2013; Crippa et al., 2016; Aas et al., 2019). Since about 2006, also Chinese emissions declined
20 by nearly 70% by 2017 (Silver et al., 2018; Zheng et al., 2018b; Mortier et al., 2020; Tong et al., 2020) (*high*
21 *confidence*). The estimated reduction in China contrasts continuing strong growth of SO₂ emissions in South
22 Asia (Figure 6.19). In 2014, over 80% of anthropogenic SO₂ emissions originated from power plants and
23 industry with Asian sources contributing more than 50% of total (Figure 6.3).

24
25 Global emissions of NO_x have been growing in spite of the successful reduction of emissions in North
26 America, Europe, Japan, Korea (Crippa et al., 2016; Turnock et al., 2016; Miyazaki et al., 2017; Jiang et al.,
27 2018a) partly driven by continuous efforts to strengthen the emission standards for road vehicles in most
28 countries (Figure 6.18 & 6.19). In many regions, an increase in vehicle fleet as well as non-compliance with
29 emission standards (Anenberg et al., 2017, 2019; Jonson et al., 2017; Jiang et al., 2018a), growing aviation
30 (Grewe et al., 2019; Lee et al., 2021) and demand for energy, and consequently large number of new fossil
31 fuel power plants, have been overcompensating these reductions. Since about 2011, global NO_x emissions
32 appear to have stabilized or slightly declined (*medium confidence*) but the global rate of decline has been
33 underestimated in CEDS, as recent data suggest that emission reductions in China were larger than included
34 in CEDs. (Figure 6.19 and Hoesly et al. (2018)). Recent bottom up emission estimates (Zheng et al., 2018b)
35 largely confirm what has been shown in satellite data (Liu et al., 2016a; Miyazaki et al., 2017; Silver et al.,
36 2018); a strong decline of NO₂ column over Eastern China (*high confidence*) (see Section 6.3.3.1). At a
37 global level, the estimated CEDS CO emission trends are comparable to NO_x, which has been confirmed by
38 several inverse modelling studies (see Section 6.3.3.2). Transport sector (including international shipping
39 and aviation) was the largest anthropogenic source of NO_x (about 50% of total) and contributed also over
40 25% of CO emissions in 2014, and Asia represented half and North America and Europe about 20% of
41 global total NO_x and CO (Figure 6.3).

42
43 Oil production-distribution and transport sector have dominated anthropogenic NMVOC emissions for most
44 of the 20th century (Hoesly et al., 2018) and still represent a large share (Figure 6.3). Efforts to control
45 transport emissions (i.e. increasing stringency of vehicle emission limits) were largely offset by fast growth
46 of emissions from chemical industries and solvent use as well as from fossil fuel production and distribution,
47 resulting in continued growth of global anthropogenic NMVOC emissions since 1900 (Figure 6.18) (*high*
48 *confidence*). Since AR5, there is *high confidence* that motor vehicle NMVOC emissions have sharply
49 decreased in North America and Europe in the last decades (Rossabi and Helmig, 2018), i.e., by about an
50 order of magnitude in major U.S. cities since 1990 (Bishop and Haugen, 2018; McDonald et al., 2018).
51 Increasing (since 2008) oil and gas extraction activities in North America lead to a strong growth of
52 NMVOC emissions (*high confidence*) as shown by analysis of ethane columns data (Franco et al., 2016), but
53 absolute emission amounts remain uncertain (Pétron et al., 2014; Tzompa-Sosa et al., 2019). In East Asia,
54 there is *medium confidence* in a decreasing trend of motor vehicle emissions, suggested by ambient
55 measurements in Beijing since 2002 (Wang et al., 2015) and by bottom-up estimates (Zheng et al., 2018b),

1 and decrease of residential heating emissions due to declining coal and biofuel use since 2005 (Zheng et al.,
2 2018b; Li et al., 2019b). However, total anthropogenic NMVOC emissions have increased steadily in China
3 since the mid-20th century, largely due to growing importance of the solvent use and industrial sectors (Sun
4 et al., 2018; Zheng et al., 2018b; Li et al., 2019b) (*medium evidence, high agreement*). Resulting changes in
5 the NMVOC speciated emissions might be underestimated in the current regional and global inventories. For
6 example, in the U.S., a recent study suggested an emergent shift in urban NMVOC sources from
7 transportation to chemical products (i.e. household chemicals, personal care products, solvents, etc.), which
8 is not in accordance with emission inventories currently used (McDonald et al., 2018). In many European
9 regions and cities, wood burning has been increasingly used for residential heating, partly for economic
10 reasons and because it is considered CO₂-neutral (Athanasopoulou et al., 2017); in situ measurements in
11 several cities, including Paris, suggest that wood burning explains up to half the NMVOC emissions during
12 winter (Kaltsonoudis et al., 2016; Languille et al., 2020). Due to the vast heterogeneity of sources and
13 components of VOCs, uncertainty in regional emissions and trends is higher than for most other components.

14
15 Emissions of carbonaceous aerosols (BC, OC) have been steadily increasing and their emissions have almost
16 doubled since 1950 (Hoesly et al., 2018) (*medium confidence*). Before 1950, North America and Europe
17 contributed about half of the global total but successful introduction of diesel particulate filters on road
18 vehicles (Fiebig et al., 2014; Robinson et al., 2015; Klimont et al., 2017a) and declining reliance on solid
19 fuels for heating brought in large reductions (Figure 6.19) (*high confidence*). Currently, global carbonaceous
20 aerosol emissions originate primarily from Asia and Africa (Bond et al., 2013; Hoesly et al., 2018; Elguindi
21 et al., 2020; McDuffie et al., 2020), representing about 80% of global total (Figure 6.3) (*high confidence*).
22 Consideration, in CMIP6, of emissions from kerosene lamps and gas flaring, revised estimates for open
23 burning of waste, regional coal consumption, and new estimates for Russia (Stohl et al., 2013; Huang et al.,
24 2015; Huang and Fu, 2016; Kholod et al., 2016; Conrad and Johnson, 2017; Evans et al., 2017; Klimont et
25 al., 2017a) resulted in over 15% higher global emissions of OC and BC than in the CMIP5 estimates for the
26 first decade of the 21st Century (Figure 6.18). The continued increase of BC emissions over East Asia after
27 2005, estimated in CMIP6 (Figure 6.19), has been, however, questioned recently as a steady decline of BC
28 concentrations was measured in the air masses flowing out from east coast of China (Kanaya et al., 2020),
29 which has been also estimated in recent regional bottom-up and top-down inventories (Zheng et al., 2018a;
30 Elguindi et al., 2020; McDuffie et al., 2020). Since AR5, confidence in emission estimates and trends in
31 North America and Europe has increased but high uncertainties remain for Asia and Africa that dominate
32 global emissions. Size distribution of emitted species, of importance for climate and health impacts, remains
33 uncertain and CEDS inventory does not provide such information. Overall, a factor two uncertainty in global
34 estimates of BC and OC emissions remains, with post 2005 emissions overestimated in Asia (*high*
35 *confidence*) and Africa (*medium confidence*).

36
37 Bottom up global emission estimates of CH₄ (Lamarque et al., 2010; Hoesly et al., 2018; Janssens-Maenhout
38 et al., 2019; Höglund-Isaksson et al., 2020) for the last two decades are higher than top down assessments
39 (e.g., Saunio et al., 2016, 2020) but trends from the two methods are similar and indicate continued growth
40 (*high confidence*). Larger discrepancies exist at the sectoral and regional levels, notably for coal mining
41 (Peng et al., 2016; Miller et al., 2019) and oil and gas sector due to growth of unconventional production and
42 higher loss estimates (Franco et al., 2016; Alvarez et al., 2018; Dalsøren et al., 2018) (see also Section 5.2.2).

43
44 Agricultural production (livestock and mineral nitrogen fertilizer application) is the primarily source of
45 ammonia in the atmosphere with more than half of present day emissions originating in Asia (Hoesly et al.,
46 2018; Figure 6.3, EC-JRC / PBL, 2020; Vira et al., 2020). NH₃ emissions are estimated to have grown
47 strongly since 1850, especially since 1950 driven by continuously increasing livestock production,
48 widespread application of mineral nitrogen fertilizers and lack of action to control ammonia (Erisman et al.,
49 2008; Riddick et al., 2016; Hoesly et al., 2018; Fowler et al., 2020) (*high confidence*). The trends estimated
50 in CMIP5 and CMIP6 are similar, while in absolute terms CMIP6 has somewhat higher emissions as it
51 includes emissions from wastewater and human waste that were largely missing in CMIP5 (Hoesly et al.,
52 2018). CMIP6 has improved spatial and temporal distribution of emissions (e.g., Lamarque et al., 2013a)
53 relying on the EDGAR v4.3 and Paulot et al. (2014), but important uncertainties remain for regionally-
54 specific temporal patterns (Riddick et al., 2016; Liu et al., 2019; Feng et al., 2020; Vira et al., 2020). The
55 continuing increase in global NH₃ emissions is driven primarily by growing livestock and crop production in

1 Asia while emissions in the US and Europe remain about constant or slightly decline in the last decade
2 (Hoesly et al., 2018). Recent satellite and ground observations support trends estimated in CMIP6 dataset
3 (Section 6.3.3.4).

4
5 To summarize, there are significant differences in spatial and temporal emission patterns of SLCFs across
6 global regions (Figure 6.18). Until the 1950s, the majority of SLCF emissions associated with fossil fuel use
7 (SO₂, NO_x, NMVOCs, CO) and about half of BC and OC originated from North America and Europe
8 (Lamarque et al., 2010; Hoesly et al., 2018). Since the 1990s a large redistribution of emission was
9 associated with strong economic growth in Asia and declining emissions in North America and Europe due
10 to air quality legislation and declining capacity of energy intensive industry, and currently more than 50% of
11 anthropogenic emissions of each SLCF species (including CH₄ and NH₃) originates from Asia (Amann et al.,
12 2013; Bond et al., 2013; Fiore et al., 2015; Crippa et al., 2016, 2018; Klimont et al., 2017a; Hoesly et al.,
13 2018) (Figure 6.3). The dominance of Asia for SLCF emissions is corroborated by growing remote sensing
14 capacity that has been providing independent evaluation of estimated pollution trends in the last decade
15 (Duncan et al., 2013; Lamsal et al., 2015; Luo et al., 2015; Fioletov et al., 2016; Geddes et al., 2016; Irie et
16 al., 2016; Krotkov et al., 2016; Wen et al., 2018).

17
18 Since AR5, the quality and completeness of activity and emission factor data and applied methodology,
19 including spatial allocation together with independent satellite-derived observations, have improved, raising
20 confidence in methods used to derive emissions. There is *high confidence* in the sign of global trends of
21 SLCF emissions until the year 2000. However, only *medium confidence* for the rate of change in the two last
22 decades owing primarily to uncertainties in actual application of reduction technologies in fast growing
23 economies of Asia. At a regional level, bottom-up derived SLCF emission trends and magnitudes in regions
24 with strong economic growth and changing air quality regulation are highly uncertain and better constrained
25 with top-down methods (Section 6.3). For most SLCF species, there is *high confidence* in trends and
26 magnitudes for affluent OECD regions where accurate and detailed information about drivers of emissions
27 exists; *medium confidence* is assessed for regional emissions of NH₃, CH₄ and NMVOC.

28
29
30 **[START FIGURE 6.3 HERE]**

31
32 **Figure 6.3: Relative regional and sectoral contributions to the present day (year 2014) anthropogenic emissions**
33 **of Short Lived Climate Forcers (SLCFs).** Emission data are from the Community Emissions Data
34 System (CEDs) (Hoesly et al., 2018). Emissions are aggregated into the following sectors: fossil fuel
35 production and distribution (coal mining, oil and gas production, upstream gas flaring, gas distribution
36 networks), fossil fuel combustion for energy (power plants), residential and commercial (fossil and
37 biofuel use for cooking and heating), industry (combustion and production processes, solvent use losses
38 from production and end use), land transportation (road and off-road vehicles), shipping (including
39 international shipping), aviation (including international aviation), agriculture (livestock and crop
40 production), waste management (solid waste, including landfills and open trash burning, residential and
41 industrial waste water), and other. Further details on data sources and processing are available in the
42 chapter data table (Table 6.SM.1).

43
44 **[END FIGURE 6.3 HERE]**

45 46 47 **6.2.2 Emissions by natural systems**

48
49 This section assesses our current understanding of SLCF emissions by natural systems. Many naturally
50 occurring emission processes in the Earth System have been perturbed by the growing influence of human
51 activities either directly (e.g., deforestation, agriculture) or via human-induced atmospheric CO₂ increase and
52 climate change and can therefore not be considered as purely natural emissions. The temporal evolution and
53 spatial distribution of natural SLCF emissions are highly variable and their estimates rely on models with
54 rather uncertain parameterizations for production mechanisms. For some SLCFs, the natural processes by
55 which emissions occur are also not well-understood. In the following sections, we assess the level of
56 confidence in present-day SLCF emissions by natural systems, in their perturbation since preindustrial and

1 their sensitivity to future changes. When available, the assessment also includes estimates from CMIP6
2 model ensemble. Note that volcanic SO₂ emissions are discussed in Section 2.2.2.

3 4 5 6.2.2.1 *Lightning NO_x*

6
7 Lightning contributes ~10% of the total NO_x emissions (Murray, 2016). Since lightning NO_x (LNO_x) is
8 predominantly released in the upper troposphere, it has a disproportionately large impact on O₃ and OH, and
9 on the lifetime of CH₄ compared with surface NO_x emissions. Whereas the global spatial and temporal
10 distribution of lightning flashes can be characterized thanks to satellite-borne and ground sensors
11 (DuplicateVirts, 2013; Cecil et al., 2014), constraining the amount of NO_x produced per flash (Miyazaki et
12 al., 2014; Medici et al., 2017; Nault et al., 2017; Marais et al., 2018; Allen et al., 2019a; Bucsele et al., 2019)
13 and its vertical allocation (Koshak et al., 2014; Medici et al., 2017) has been more elusive. Atmospheric
14 Chemistry and Climate Model Intercomparison Project (ACCMIP) models in CMIP5 used a range of LNO_x
15 between 1.2 to 9.7 TgN year⁻¹ (Lamarque et al., 2013c). In CMIP6, the corresponding LNO_x range is
16 between 3.2 to 7.6 TgN year⁻¹ (Griffiths et al., 2020). All CMIP6 models (as well as most models included in
17 the CMIP5, Young et al., 2013) apply a parameterization that relates cloud-top-height to lightning intensity
18 (Price and Rind, 1992), projecting an increase in LNO_x in a warmer world in the range 0.27 to 0.61 Tg(N)
19 yr⁻¹ per °C (Thornhill et al., 2021a). However, models using parameterizations based on convection (Grewe
20 et al., 2001), updraft mass flux (Allen and Pickering, 2002), or ice flux (Finney et al., 2016a) show either
21 much less sensitivity or a negative response (Finney et al., 2016b, 2018; Clark et al., 2017).
22 In summary, the total present-day global lightning NO_x emissions are still estimated to be within a factor of
23 2. There is *high confidence* that LNO_x are perturbed by climate change however there is *low confidence* in
24 the sign of the change due to fundamental uncertainties in parameterizations.

25 26 27 6.2.2.2 *NO_x emissions by soils*

28
29 Soil NO_x (SNO_x) emissions occur in connection with complex biogenic/microbial nitrification and
30 denitrification processes (Ciais et al., 2013), which in turn are sensitive – in a non-linear manner – to
31 temperature, precipitation, soil moisture, carbon and nutrient content and the biome itself (e.g., Hudman et
32 al., 2012). Global SNO_x estimates, based on observationally constrained chemistry-transport model and
33 vegetation model studies, show a broad range between 4.7 and 16.8 TgN/yr (Young et al., 2018). This
34 estimate is generally larger than the current source strength used in CMIP6 simulations, which is prescribed
35 using an early empirical estimate, typically scaled to about 5 TgN/yr (Yienger and Levy, 1995)
36 Under warmer climate, the overall nitrogen fixation in non-agricultural ecosystems is expected to be 40%
37 larger than in 2000, due to increased enzyme activity with growing temperatures, but the emission rates of
38 NO (and N₂O) is expected to be dominated by changes in precipitation patterns and evapotranspiration fluxes
39 (Fowler et al., 2015). Current earth system models (ESM) incorporate biophysical and biogeochemical
40 processes only to a limited extent (Jia et al., 2019), precluding adequate climate sensitivity studies for SNO_x.
41 Hence, while the current strength source of soil NO_x has been better constrained over the last decade,
42 adequate representations of SNO_x and how it escapes from the canopy are still missing in ESM to provide
43 quantitative estimates of climate driven changes in SNO_x.

44 45 6.2.2.3 *Vegetation emissions of organic compounds*

46
47 A wide range of BVOCs are emitted from vegetation with the dominant compounds being isoprene and
48 monoterpenes but also including sesquiterpenes, alkenes, alcohols, aldehydes and ketones. The photo-
49 oxidation of BVOC emissions plays a fundamental role in atmospheric composition by controlling the
50 regional and global budgets of ozone and organic aerosols, and impacting the lifetime of methane and other
51 reactive components (Arneth et al., 2010b; Heald and Spracklen, 2015). Substantial uncertainty exists across
52 different modelling frameworks for estimates of global total BVOC emissions and individual compound
53 emissions (Messina et al., 2016). Global isoprene emission estimates differ by a factor of 2 from 300-600
54 Tg(C) yr⁻¹ and global monoterpene emission estimates by a factor of 5 from 30-150 Tg(C) yr⁻¹ (Messina et
55 al., 2016). A main driver of the uncertainty ranges is the choice of assignment of PFT-specific basal emission

1 rates in the model, however, the smaller uncertainty range for isoprene than for monoterpenes is not fully
2 understood (Arneth et al., 2008). The evaluation of global BVOC emissions is challenging because of poor
3 measurement data coverage in many regions and the lack of year-round measurements (Unger et al., 2013).
4 Several observational approaches have been developed in the past few years to improve understanding of
5 BVOC emissions including indirect methods such as the measurement of the OH loss rate in forested
6 environments (Yang et al., 2016) and application of the variability in satellite formaldehyde concentrations
7 (Palmer et al., 2006; Barkley et al., 2013; Stavrakou et al., 2014). Direct space-borne isoprene retrievals
8 using infrared (IR) radiance measurements have very recently become available (Fu et al., 2019; Wells et al.,
9 2020). Collectively these approaches have identified weaknesses in the ability of the parameterizations in
10 global models to reproduce BVOC emissions hotspots (Wells et al., 2020). However, none of the current
11 observational approaches have yet been able to reduce the uncertainty ranges in global emission estimates.

12
13 At the plant level, BVOC emission rates and composition depend strongly on plant species with plants
14 tending to emit either isoprene or monoterpenes but not both. Photosynthetic activity is a main driver of
15 isoprene and monoterpene production. Therefore, radiation and temperature, along with leaf water status,
16 phenological state and atmospheric CO₂ mixing ratio, affect emissions directly (on the leaf-scale) and
17 indirectly (via plant productivity) (Guenther et al., 2012; Loreto et al., 2014; Niinemets et al., 2014). CO₂
18 directly influences the isoprene synthesis process, with inhibition under increasing atmospheric CO₂
19 (Rosenstiel et al., 2003; Possell et al., 2005; Wilkinson et al., 2009). Direct CO₂ inhibition has been observed
20 for some monoterpene compounds (Loreto et al., 2001; Llorens et al., 2009). Severe/long-term water stress
21 may reduce emissions whilst mild/short-term water stress may temporarily amplify or maintain BVOC
22 emissions to protect plants against on-going stress (Peñuelas and Staudt, 2010; Potosnak et al., 2014;
23 Genard-Zielinski et al., 2018). Furthermore, observations in the Amazon, indicate the chemical composition
24 of monoterpene emissions could also change under elevated temperature conditions (Jardine et al., 2016). In
25 addition, all these processes are investigated over short-time scales but the long-term response of BVOC
26 emissions depends on how the vegetation itself responds to the altered climate state (including temperature
27 and water stress).

28
29 Global BVOC emissions are highly sensitive to environmental change including changes in climate,
30 atmospheric CO₂ and vegetation composition and cover changes in natural and managed lands. Recent global
31 modelling studies agree that global isoprene emissions have declined since the preindustrial driven
32 predominantly by anthropogenic LULCC with results converging on a 10-25% loss of isoprene emissions
33 globally between 1850 and present day (Lathière et al., 2010; Unger, 2013, 2014; Acosta Navarro et al.,
34 2014; Heald and Geddes, 2016a; Hantson et al., 2017; Hollaway et al., 2017; Scott et al., 2017). The
35 historical evolution of monoterpene and sesquiterpene emissions is less well studied and there is no robust
36 consensus on even the sign of the change (Acosta Navarro et al., 2014; Hantson et al., 2017). Future global
37 isoprene and monoterpene emissions depend strongly on the climate and land-use scenarios considered
38 (Hantson et al., 2017; Szogs et al., 2017). BVOC emissions will be sensitive to future land-based climate
39 change mitigation strategies including afforestation and bioenergy, with impacts of bioenergy depending on
40 the choice of crops (Szogs et al., 2017).

41
42 Most CMIP6 models use overly simplistic parameterizations and project an increase in global BVOC
43 emissions in response to warming temperatures (Turnock et al., 2020). This good agreement actually reflects
44 the lack of diversity in BVOC emission parametrizations in global models that do not fully account for the
45 complex processes influencing emissions discussed above.

46
47 Overall, we assess that historical global isoprene emissions declined between the preindustrial and present
48 day by 10-25% (*low confidence*) but historical changes in global monoterpenes and sesquiterpenes are too
49 uncertain to provide an assessment. Future changes in BVOCs depend strongly on the evolution of climate
50 and land-use and are strongly sensitive to land-based climate change mitigation strategies. However, the net
51 response of BVOC emissions is uncertain due to complexity of processes, hard to constrain observationally,
52 considered with various degrees of details in models.

6.2.2.4 Land emissions of dust particles

The emission of dust particles into the atmosphere results from a natural process, namely saltation bombardment of the soil by large wind-blown particles such as sand grains and from disintegration of saltating particle clusters (Kok et al., 2012). The occurrence and intensity of dust emissions are controlled by soil properties, vegetation, and near-surface wind, making dust emissions sensitive to changes in climate, and land-use and land cover (Jia et al., 2019). In addition, dust can be directly emitted through human activities, such as agriculture, off-road vehicles, building construction, mining, and indirectly emitted through hydrological changes due to human actions such water diversion for irrigation (e.g. Ginoux et al., 2012). Estimates of the anthropogenic fraction of global dust vary from less than 10% to over 60% suggesting that human contribution to the global dust budget is quite uncertain (Ginoux et al., 2012; Stanelle et al., 2014; Xi and Sokolik, 2016). Reconstruction of global dust (deposition) from paleo records indicate factor of 2 to 4 changes between the different climate regimes in the glacial and interglacial periods (see Section 2.2.6). An extremely limited number of studies have explored the evolution of global dust sources since preindustrial times (Mahowald et al., 2010; Stanelle et al., 2014). A modelling study estimated a 25% increase in global dust emissions between the late nineteenth century to present due to agricultural land expansion and climate change (Stanelle et al., 2014). CMIP5 models were unable to capture the observed variability of annual and longer time scales in North African dust emissions (Evan et al., 2014), however more recent ESMs with process-based dust emission schemes that account for changes in vegetation and climate in a more consistent manner better match the observations (Kok et al., 2014; Evans et al., 2016). Feedbacks between the global dust cycle and the climate system (see Section 6.4.6) could account for a substantial fraction of the total aerosol feedbacks in the climate system with an order of magnitude enhancement on regional scale (Kok et al., 2018). In summary, there is *high confidence* that atmospheric dust source and loading are sensitive to changes in climate and land use, however, there is *low confidence* in quantitative estimates of dust emission response to climate change.

6.2.2.5 Oceanic emissions of marine aerosols and precursors

Oceans are a significant source of marine aerosols that influence climate directly by scattering and absorbing solar radiation or indirectly through the formation of cloud condensation nuclei (CCN) and ice nucleating particles (INP). Marine aerosols consist of primary sea spray particles and secondary aerosols produced by the oxidation of emitted precursors, such as dimethylsulphide (DMS) and numerous other BVOCs. Sea spray particles, composed of sea-salt and primary organic aerosols (POA), are produced by wind induced wave breaking as well as direct mechanical disruption of waves. The understanding of sea spray emissions has increased substantially over the last five years, however, the knowledge of formation pathways and factors influencing their emissions continue to have large uncertainties (Forestieri et al., 2018; Saliba et al., 2019). The emission rate of sea-spray particles is predominantly controlled by wind speed. Since AR5, the influence of other factors, including sea surface temperature, wave history and salinity is increasingly evident (Callaghan et al., 2014; Grythe et al., 2014; Ovadnevaite et al., 2014; Salter et al., 2014; Barthel et al., 2019). Marine POA, often the dominant submicron component of sea spray, are emitted as a result of oceanic biological activity, however the biological processes by which these particles are produced remain poorly characterized contributing to large uncertainties in global marine POA emission estimates (Tsigaridis et al., 2014; Burrows et al., 2018; Cravigan et al., 2020; Hodzic et al., 2020). Furthermore, particle size and chemical composition of sea spray particles and how these evolve in response to changing climate factors and dynamic oceanic biology continue to have large uncertainties.

DMS, the largest natural source of sulphur in the atmosphere, is produced by marine phytoplankton and is transferred from ocean water to the atmosphere due to wind-induced mixing of surface water. DMS oxidizes to produce sulphate aerosols and contributes to the formation of CCN. Since AR5, the range in global DMS flux estimates reduced from 10-40 Tg S year⁻¹ to 9 to 34 Tg S year⁻¹ with a *very likely* range of 18-24 Tg S year⁻¹ based on sea-surface measurements and satellite observations (Lana et al., 2011). DMS production, and consequently emissions, have been shown to respond to multiple stressors, including climate warming, eutrophication, and ocean acidification. However, large uncertainties in process-based understanding of the mechanisms controlling DMS emissions, from physiological to ecological, limit our knowledge of past

1 variations and our capacity to predict future changes.

2
3 Overall, there is *low confidence* in the magnitude and changes in marine aerosol emissions in response to
4 shifts in climate and marine ecosystem processes.

5 6 7 6.2.2.6 *Open biomass burning emissions*

8
9 Open biomass burning (including forest, grassland, peat fires and agricultural waste burning) represent about
10 30%, 10%, 15% and 40% of present-day global emissions of CO, NO_x, BC, and OC, respectively (van Marle
11 et al., 2017; Hoesly et al., 2018). Wildfires also play an important role in several atmospheric chemistry–
12 climate feedback mechanisms (Bowman et al., 2009; Fiore et al., 2012) and fire events occurring near
13 populated areas induce severe air pollution episodes (Marlier et al., 2020; Rooney et al., 2020; Yu et al.,
14 2020).

15
16 For the last two decades, model-based emission estimates are constrained by remote sensing capacity to
17 detect active fires and area burned. In AR5, biomass burning emissions were derived from a satellite product
18 (Lamarque et al., 2010). Since then, improvements in detection of small fires has enhanced the agreement
19 with higher-resolution and ground-based data on burned area in several regions (Randerson et al., 2012;
20 Mangeon et al., 2015) especially for areas subjected to agricultural waste burning (Chuvienco et al., 2016,
21 2019). The updated emission factors and contribution of forest versus savannah fires lead to significantly
22 higher global emissions of NO_x and lower emissions of OC and CO in CMIP6, compared with CMIP5. A
23 recent compilation and assessment of emission factors (Andreae, 2019) indicates that the emission factors
24 from Akagi et al. (2011), primarily used to produce the CMIP6 datasets, differ within ±50% for CO, OC,
25 BC, and NO_x, depending on the biome, and would imply, for example, up to 10-30% higher OC and BC
26 emissions from tropical forest fires.

27
28 The historical (pre-satellite era) dataset for CMIP6 considers advances in knowledge of past fire dynamics
29 (new fire proxy datasets, such as charcoal in sediments or levoglucosan in ice-cores) and visibility records
30 from weather stations (Marlon et al., 2016; van Marle et al., 2017). At a global level, CMIP5 and CMIP6
31 emission trends are similar, however, there are substantial differences at the regional level, especially for the
32 United States, South America (south of Amazonia), and southern hemisphere Africa (van Marle et al., 2017).

33
34 Globally, the CMIP5 estimates (Lamarque et al., 2010), indicated gradual decline of open biomass burning
35 emission from 1920 to about 1950 and then steady, and stronger than CMIP6, increase towards 2000. In
36 contrast, CMIP6 biomass burning emissions (van Marle et al., 2017) increase only slightly over 1750 to 2015
37 and peak during the 1990s after which they decrease gradually consistently, with the assessment of fire
38 trends in Chapter 5. Therefore, the CMIP6 evolution has a smaller difference between pre-industrial and
39 present emissions than CMIP5, resulting in a lower radiative forcing of biomass burning SLCFs, leading to
40 possibly lower effect on climate (van Marle et al., 2017).

41
42 Climate warming, especially through change in temperature and precipitation, will generally increase the risk
43 of fire (Jia et al., 2019, see also Chapter 12) and can also affect the fire injection and plume height (Veira et
44 al., 2016), but occurrence of fires and their emissions in the future strongly depends on anthropogenic factors
45 such as population density, land use and fire management (Veira et al., 2016). Consequently, future
46 emissions vary widely with increases and decreases amongst the SSP scenarios due to different land use
47 change scenarios.

48
49 In summary, there has been an improvement in the knowledge of biomass burning emissions by reducing
50 key uncertainties highlighted in AR5. However, systematic assessment of remaining uncertainties is limited,
51 with a lower limit of uncertainties due to emission factors of 30 %, and larger uncertainties due to burning
52 activity estimates, especially at regional level. Overall, a *medium confidence* in current global biomass
53 burning SLCF emissions and their evolution over the satellite era is assessed. There is *low-to-medium*
54 *confidence* in pre-industrial to the 1980s of biomass SLCF emissions, which rely on incorporation of several
55 proxy data, with limited spatial representativeness. Nevertheless, uncertainties in absolute value of pre-

1 industrial emissions remain high, limiting confidence of radiative forcing estimates.
2
3

4 **6.3 Evolution of Atmospheric SLCF abundances** 5

6 This section assesses the evolution of atmospheric abundance¹ of SLCFs since AR5 based on observations
7 and modelling, our knowledge of SLCFs burden and distribution, and our understanding of the trends over
8 longer time scales. In addition to emissions (section 6.2), atmospheric chemistry (gas and aqueous
9 chemistry), deposition (including wet and dry removal), and transport processes play a major role in
10 determining the atmospheric distribution, budget and lifetime of SLCFs. The distribution and lifetime of
11 SLCFs are further influenced by the modulation of chemical and physical processes in response to a
12 changing climate. Therefore, the time evolution of atmospheric abundance of SLCFs is characterized by
13 many complex non-linear interactions occurring at varying temporal and spatial scales. For this assessment,
14 global scale, long-term measurements are employed only for a few gaseous SLCFs while for most short-
15 lived species regional-scale observations and global models are relied upon.
16

17
18 **[START BOX 6.1 HERE]**
19

20 **BOX 6.1: Atmospheric abundance of SLCFs: from process level studies to global chemistry-climate 21 models** 22

23 Changes in the atmospheric distribution of SLCFs determine their radiative forcing, and climate and air
24 quality impacts. This box provides an overview of how process level understanding of the distribution and
25 evolution of chemical compounds is derived and where uncertainties come from.
26

27 Process-level understanding of tropospheric gas and aerosol chemistry developed through laboratory and
28 simulation chamber experiments as well as quantum chemical theory is used to generate chemical
29 mechanisms. Atmospheric simulation chambers are designed to identify the chemical pathways and quantify
30 reaction kinetics in isolation from atmospheric transport, deposition and emission processes. Ideally the
31 chemical regimes studied, are representative for ambient atmospheric complexity and concentrations (e.g.,
32 McFiggans et al., 2019). Recently, quantum chemical theory has advanced to a level that it can provide
33 kinetic and product information in a parameter range not possible with laboratory experiments (Vereecken et
34 al., 2015). Iterative and interlinked use of simulation chamber and quantum chemical theory has led to
35 improved knowledge of chemical mechanisms (Fuchs et al., 2013a; Nguyen et al., 2010; Peeters et al., 2009,
36 2014). For application in chemistry climate models (CCMs), the chemical mechanisms need to be
37 computationally efficient, requiring simplifications. Such simplifications include reduced hydrocarbon
38 representations, the application of lumping techniques (one compound or a chemical structure representing a
39 family of compounds, for example, as done for parameterizing SOA formation) and/or the implementation of
40 artificial operators representing key steps of the chemistry (Emmerson and Evans, 2009; Xia et al., 2009;
41 Stockwell et al., 2020). Additionally, aerosol microphysical processes (nucleation, coagulation,
42 condensation, evaporation, and sedimentation) that determine the evolution of aerosol number concentrations
43 and size particle distribution are represented in parameterised forms in global models with varying levels of
44 complexity (Mann et al., 2014).
45

46 A wide range of in situ and remotely sensed observations are used to characterise atmospheric chemical
47 composition. Measurements made routinely as part of long-term monitoring programmes are particularly
48 useful for assessing long-term trends and variability, and spatial distributions (as in Section 2.2, Section 6.3,
49 and Section 7.3.3) while intensive field campaigns provide a more comprehensive view of atmospheric
50 composition at a specific location for a limited time facilitating an improved process-level understanding.
51 Retrieval of atmospheric concentrations from satellites, in particular, has been tremendously useful for
52 providing global continuous coverage, although the retrievals themselves depend on prior information of
53 atmospheric composition usually derived from models. Over the last decade or so, observations of

¹ The word 'concentration' is used to denote abundances in terms of mixing ratio for most species unless specified

1 atmospheric concentrations have been combined with information from global chemistry-climate models to
2 produce global assimilation and forecasting systems with the purpose of producing chemical reanalysis or
3 improving model inputs (i.e., emissions or boundary conditions) and forecasts (Miyazaki et al., 2015;
4 Randles et al., 2017; Inness et al., 2019).

5
6 Global three-dimensional CCMs (Box 6.1, Figure 1) represent the full coupling of chemistry with climate
7 physics (e.g., Morgenstern et al. 2017) with different levels of complexity (e.g. interactive aerosols with or
8 without tropospheric and/or stratospheric chemistry). Methane concentrations are typically prescribed or
9 constrained to observations while emissions of other SLCFs (or their precursors) are either prescribed or
10 calculated interactively in the current generation of CCMs (Collins et al., 2017). CCMs, now part of Earth
11 System Models (ESMs), are applied extensively to simulate the distribution and evolution of chemical
12 compounds on a variety of spatial and temporal scales to improve current knowledge, make future
13 projections and investigate global scale chemistry-climate interactions and feedbacks (see also Chapter 3
14 Section 3.8.2.2). CCMs are also used to interpret observations to disentangle the processes that drive
15 observed variability and trends. Some aspects of air quality, such as diurnal peaks or local threshold
16 violations, strong gradient in chemical regimes and coupling between processes cannot be captured by
17 relatively coarse spatial resolution (> 50 km) global CCMs (Markakis et al., 2014) and necessitate
18 subsequent downscaling modelling exercises.

19
20 The skill of CCMs is typically assessed by their ability to reproduce observed abundance, and trends and
21 variability of chemical compounds. However, uncertainty remains large because of observation limitations
22 (errors and uncertainties, spatial and temporal coverage), model parameterizations (e.g., chemical
23 mechanisms, photolysis schemes, parameterizations for mixing and convective transport, and deposition),
24 model input parameters (e.g., reaction rate constants, emissions) and an incomplete understanding of the
25 physical and chemical processes that determine SLCF distributions (Brasseur and Jacob, 2017; Young et al.,
26 2018). CCMs can therefore not capture every aspect of atmospheric chemical composition, but are expected
27 to represent, as faithfully as possible, the sensitivity of chemical compounds to their drivers (e.g.,
28 anthropogenic emissions). Models are evaluated in multiple ways to identify their strengths and weaknesses
29 in explaining the evolution of SLCF abundances. For example, CCM simulations are performed in the
30 nudged or offline meteorology mode, that is, driven by observed or reanalysed meteorology rather than in the
31 free-running mode, for consistent comparison of modelled chemical composition with observations for a
32 specific time period (Dameris and Jöckel, 2013). Although, caution is exercised as nudging can alter the
33 model climate resulting in unintentional impacts on the simulated atmospheric physics and/or chemistry
34 (Orbe et al., 2018; Chrysanthou et al., 2019). Chemical mechanisms implemented in CCMs are evaluated
35 and intercompared to assess their skill in capturing relevant chemistry features (e.g., Brown-Steiner et al.
36 2018). The multi-model ensemble approach, employed for evaluating climate models, has been particularly
37 useful for characterizing errors in CCM simulations of SLCFs related to structural uncertainty and internal
38 variability (Naik et al., 2013b; Shindell et al., 2013; Turnock et al., 2020; Young et al., 2013). However, as
39 discussed in BOX 4.1, this approach is unable to capture the full uncertainty range.

40
41 This assessment draws upon results from single model studies and recent multi-model intercomparisons
42 (e.g., AeroCom, CCMI), in particular those endorsed by CMIP6, which then allows for the full consideration
43 of robustness and uncertainty due to model structures and processes. Based on the collective information
44 provided in this body of literature, the CMIP6 multimodel ensemble is largely fit-for-purpose of evaluating
45 the influence of SLCFs on radiative forcing, climate and non-CO₂ biogeochemical feedbacks. Additionally,
46 CMIP6 models are fit for capturing global air pollution response to changes in emissions and meteorology
47 but have difficulty in simulating the mean state (Turnock et al., 2020). The set of CMIP6 simulations has
48 been used to update the relations between emissions and surface temperature at the heart of the emulators
49 (Cross-Chapter box 7.1) and update emission metrics (Section 7.6). Emulators and emission metrics are used
50 in this chapter (Sections 6.5 and 6.7) to assess more specifically the effect of the individual SLCFs for each
51 sector and region which would be of prohibitive computing cost with CCMs. CCMs are also used to build
52 global source-receptor models which use relations between surface concentrations and emissions. Such a
53 model is used to assess the impact of various mitigation on air quality (Sections 6.5 and 6.7.)
54
55

1 [START BOX 6.1, FIGURE 1 HERE]

2
3 **Box 6.1, Figure 1: Knowledge exchange between laboratory/theoretical studies, observations and global**
4 **chemistry-climate models to inform our understanding of short-lived climate forcers (SLCFs).**

5
6 [END BOX 6.1, FIGURE 1 HERE]

7
8 [END BOX 6.1 HERE]

9
10
11 **6.3.1 Methane (CH₄)**

12
13 The global mean surface mixing ratio of methane has increased by 156% since 1750 (see Section 2.2.3.4;
14 Annex III). Since AR5, the methane mixing ratio has increased by about 3.5% from 1803±2 ppb in 2011 to
15 1866±3 ppb in 2019 (see Section 2.2.3.3.2) largely driven by anthropogenic activities as assessed in Chapter
16 5 (Section 5.2.2 and Cross-Chapter Box 5.2).

17
18 An assessment of the global methane budget is provided in Chapter 5, while this section assesses methane
19 atmospheric lifetime and perturbation time (Prather et al., 2001). AR5 based its assessment of methane
20 lifetime on (Prather et al., 2012). The methane lifetime due to tropospheric OH, the primary sink of methane,
21 was assessed to be 11.2 ± 1.3 years constrained by surface observations of methyl chloroform (MCF), and
22 lifetimes due to stratospheric loss², tropospheric halogen loss and soil uptake were assessed to be 150 ± 50
23 years, 200 ± 100 years, 120 ± 24 years, respectively (Myhre et al., 2013b). Considering the full range of
24 individual lifetimes, the total methane lifetime was assessed in AR5 to be 9.25 ± 0.6 years.

25
26 The global chemical methane sink, essentially due to tropospheric OH, required to calculate the chemical
27 lifetime is estimated by either bottom-up global CCMs and ESMs (BU) or top-down observational inversion
28 methods (TD). BU global models represent the coupled chemical processes and feedbacks that determine the
29 chemical sinks but show large diversity in their estimates, particularly the tropospheric OH sink (Zhao et al.,
30 2019b; Stevenson et al., 2020). TD inversion methods, on the contrary, provide independent observational
31 constraints on methane sink due to tropospheric OH over large spatio-temporal scales, but are prone to
32 observational uncertainties and do not account for the chemical feedbacks on OH (Prather and Holmes,
33 2017; Naus et al., 2019). The central estimate of mean chemical methane loss over the period 2008-2017
34 varies from 602 [minimum and maximum range of 507-803] Tg yr⁻¹ from BU chemistry-climate models in
35 the Chemistry Climate Modeling Initiative (CCMI) to 514 [474-529] Tg yr⁻¹ from TD inverse modelling
36 (also see section 5.2.2 and Table 5.2). The smaller range in TD estimate (11%) results from the use of a
37 common climatological mean OH distribution (Saunio et al., 2020; Zhao et al., 2020a) while the larger
38 range in BU estimate (49%) reflects the diversity in OH concentrations from different chemical mechanisms
39 implemented in the global models (Zhao et al., 2019b). See Section 6.3.3 for further discussion on the
40 conflicting information on OH from CCMs/ESMs and TD inversion approaches. Further work is required to
41 reconcile differences between BU and TD estimates of the chemical methane sink.

42
43
44 [START TABLE 6.2 HERE]

45
46 **Table 6.2:** Methane lifetime due to chemical losses, soil uptake, and total atmospheric lifetime based on CMIP6
47 multi-model analysis, and bottom-up and top-down methane budget estimates in Table 5.2. Bottom-up
48 and top-down methane lifetimes are calculated using the central estimates of the respective sinks for the
49 mean 2008-2017 period in Table 5.2 together with mean 2008-2017 global methane concentration of
50 1815 ppb (see Annex III) converted to methane burden using a fill-factor of 2.75 Tg/ppb from Prather et
51 al., (2012). Values in parentheses show the minimum and maximum range.

2 Prather et al (2012) report lifetime due to stratospheric loss and soil uptake as 120 ± 24 years and 150 ± 50 years, respectively, and they were inadvertently swapped in AR5 with no effect on the total methane lifetime.

Study	Total Chemical Lifetime (yr)	Soil Lifetime (yr)	Total Atmospheric Lifetime (yr)	Number of models/inversions
(Stevenson et al., 2020) ^a	8.3 (8.1-8.6) ^b	160	8.0 (7.7-8.2)	3 CMIP6 ESMs
Bottom-up (based on Table 5.2)	8.3 (6.2-9.8)	166 (102-453)	8.0 (6.3-10.0)	7 CCMI CCMs/CTMs
Top-down (based on Table 5.2)	9.7 (9.4-10.5)	135 (116-185)	9.1 (8.7-10.0)	7 inversion systems
AR6 assessed value ^c	9.7 ± 1.1	135 ± 44	9.1 ± 0.9	Based on Top-down with uncertainty estimate from AR5

^a mean over 2005-2014

^b does not include lifetime due to tropospheric halogen loss

^c uncertainties indicate ± 1-standard deviation

[END TABLE 6.2HERE]

The present-day BU methane chemical lifetime shows a larger spread than that in the TD estimates (Table 6.2) in line with the spread seen in the sink estimates. The spread in the methane lifetime calculated by three CMIP6 ESMs is narrower and is enclosed within the spread of the BU CCMI model ensemble. Based on the consideration that the small imbalance in total methane sources versus sinks derived from TD estimates is close to the observed atmospheric methane growth rate (see Table 5.2), the TD values are assessed to be the best estimates for this assessment. The relative uncertainty (1-sigma) is taken to be the same as that in AR5, i.e., 11.8%, 33% and 10% for chemical, soil and total lifetime, respectively. The central estimate of the total atmospheric methane lifetime assessed here is the same as that in the AR5.

The methane perturbation lifetime (τ_{pert}) is defined as the e-folding time it takes for methane burden to decay back to its initial value after being perturbed by a change in methane emissions. Perturbation lifetime is longer than the total atmospheric lifetime of methane as an increase in methane emissions decreases tropospheric OH which in turn decreases the lifetime and therefore the methane burden (Prather, 1994; Fuglestvedt et al., 1996; Holmes et al., 2013; Holmes, 2018). Since perturbation lifetime relates changes in emissions to changes in burden, it is used to determine the emissions metrics assessed in Chapter 7 Section 7.6. The perturbation lifetime is related to the atmospheric lifetime as $\tau_{\text{pert}} = f * \tau_{\text{total}}$ where f is the feedback factor and is calculated as $f = 1/(1-s)$, where $s = -\delta(\ln \tau_{\text{total}}) / \delta(\ln[\text{CH}_4])$ (Prather et al., 2001). Since there are no observational constraints for either τ_{pert} or f , these quantities are derived from CCMs or ESMs. AR5 used $f = 1.34 \pm 0.06$ based on a combination of multimodel (mostly CTMs and a few CCMs) estimates (Holmes et al., 2013). A recent model study explored new aspects of methane feedbacks finding that the strength of the feedback, typically treated as a constant, varies in space and time but will in all likelihood remain within 10% over the 21st century (Holmes, 2018). For this assessment, the value of f is assessed to be 1.30 ± 0.07 based on a six member ensemble of AerChemMIP ESMs (Thornhill et al., 2021b). This f value is slightly smaller but within the range of the AR5 value. This results in an overall perturbation methane lifetime of 11.8 ± 1.8 years, within the range of the AR5 value of 12.4 ± 1.4 years. The methane perturbation lifetime assessed here is used in the calculation of emission metrics in Section 7.6.

6.3.2 Ozone (O₃)

6.3.2.1 Tropospheric ozone

About ten percent of the total atmospheric ozone column resides in the troposphere. The ozone forcing on climate strongly depends on its vertical and latitudinal distribution in troposphere. The lifetime of ozone in the troposphere ranges from a few hours in polluted urban regions to up to few months in the upper troposphere. Observed tropospheric ozone concentrations range from less than 10 ppb over the tropical Pacific Ocean to as much as 100 ppb in the upper troposphere and more than 100 ppb downwind of major ozone precursor emissions regions. An ensemble of five CMIP6 models including whole atmospheric chemistry and interactive ozone has been shown to simulate consistently the present-day ozone distribution (N-S and latitudinal gradients) and its seasonal variability when compared with observations from sondes, background surface stations and satellite products (Griffiths et al., 2020). The biases, whose magnitude is similar to AR5, are lower than 15% against climatological seasonal cycle from ozonesondes with an overestimate in the Northern Hemisphere and an underestimate in the Southern Hemisphere (Griffiths et al., 2020). The CMIP6 multimodel ensemble estimate of the global mean lifetime of ozone for present day conditions is 25.5 ± 2.2 days (Griffiths et al., 2020) which is within the range of previous multi-model estimates (Stevenson et al., 2006; Young et al., 2013), indicating a *high level of confidence*.

AR5 assessed the tropospheric ozone burden to be 337 ± 23 Tg for year 2000 based on the ACCMIP ensemble of model simulations (Myhre et al., 2013b). Multiple satellites products, ozonesondes and CCMs are used to estimate tropospheric ozone burden (Table 6.3). Satellite products provide lower-bound values as they exclude regions under polar night conditions (Gaudel et al., 2018). The tropospheric ozone burden values from multi-model exercises are within the range of the observational estimates despite different definitions of the tropopause for multi-model estimates which can lead to differences of about 10% on the ozone burden model estimates (Griffiths et al., 2020). Weighted by their number of members, CMIP6 and CCMI multimodel estimates and observational estimates of tropospheric ozone burden about year 2010, lead to an assessment of the tropospheric ozone burden of 347 ± 28 Tg for 2010.

The tropospheric ozone budget is controlled by chemical production and loss, by STE, and by deposition at the Earth's surface, whose magnitude are calculated by CCMs (Table 6.3). Despite the high agreement of the model ensemble mean with observational estimates in present-day tropospheric ozone burden, the values of individual budget terms can vary widely across models in CMIP6 consistent with previous model intercomparison experiments (Young et al., 2018). Furthermore, single-model studies have shown that the halogen chemistry, which is typically neglected from model chemistry schemes in CCMs, may have a notable impact on the ozone budget, as halogens, particularly of marine origin, take part in efficient ozone loss catalytic cycles in the troposphere (Saiz-Lopez et al., 2012; Sarwar et al., 2015; Sherwen et al., 2016).

[START TABLE 6.3 HERE]

Table 6.3: Global tropospheric ozone budget terms and burden based on multi-model estimates and observations for present conditions. All uncertainties quoted as 1σ . Values of tropospheric ozone burden with asterisk indicate average over the latitudinal zone 60°N- 60°S.

Period	Burden (Tg)	Production Tg yr ⁻¹	Loss Tg yr ⁻¹	Deposition Tg yr ⁻¹	STE Tg yr ⁻¹	Number of Models / Reference
Models						
~ 2000 time slice (1995–2004)	347 ± 30	5283 ± 1798	4108 ± 486	1075 ± 514	626 ± 781	CMIP6 ^a (5 Earth System Models) (Griffiths et al., 2020)
~2010 time slice (2005–2014)	356 ± 31	5530 ± 1909	4304 ± 535	1102 ± 538	628 ± 804	
~ 2000	341 ± 31					

	(309±31)*					CCMI ^b (9 models) (Archibald et al., 2020)
2010	345±30 (314±29)*					
~ 2000	340 ± 34	4937 ± 656	4442 ± 570	996 ± 203	535 ± 161	TOAR ^c (based on 32-49 models participating in intermodel comparisons and single model studies) (Young et al., 2018)
Observations						
2010–2014	338 ± 6					TOST ^d , IASI ^e -FORLI, and IASI-SOFRID (Gaudel et al., 2018)
2010–2014	302 ± 12*					TOST, IASI-FORLI, IASI-SOFRID, OMI ^f /MLS, OMI-SAO and OMI-RAL (Gaudel et al., 2018)

^aCMIP6: Coupled Model Intercomparison Project Phase 6; ^bCCMI: Chemistry-Climate Model Initiative; ^c TOAR Tropospheric Ozone Assessment Report; ^d TOST Trajectory-mapped Ozone-sonde dataset for the Stratosphere and Troposphere; ^e IASI Infrared Atmospheric Sounding Interferometer; ^f OMI Ozone Monitoring Instrument.

[END TABLE 6.3 HERE]

Because of the heterogeneous distribution of ozone, limited observations or proxies do not provide accurate information about the global preindustrial abundance, posing a challenge to the estimation of the historical evolution of tropospheric ozone. Therefore, global CCMs complemented by observations are relied upon for estimating the long-term changes in tropospheric ozone. AR5 concluded that anthropogenic changes in ozone precursor emissions are unequivocally responsible for the increase between 1850 and present (Myhre et al., 2013b). Based on limited isotopic evidence, Chapter 2 assesses that the global tropospheric ozone increased by less than 40% between 1850 and 2005 (*low confidence*) (see Section 2.2.5.3). The CMIP6 models are in line with this increase of tropospheric ozone with an ensemble mean value of 109 ± 25 Tg (model range) from 1850-1859 to 2005-2014 (Figure 6.4). This increase is higher than the AR5 value of 100 ± 25 Tg from 1850 to 2010 due to higher ozone precursor emissions in CMIP6. However, the AR5 and CMIP6 value are close when considering the reported uncertainties. The uncertainties are equivalent in CMIP6 and AR5 despite enhanced inclusion of coupled processes in the CMIP6 ESMs (e.g. biogenic VOC emissions or interactive stratospheric ozone chemistry).

[START FIGURE 6.4 HERE]

Figure 6.4: Time evolution of global annual mean tropospheric ozone burden (in Tg) from 1850 to 2100. Multi-model means for CMIP6 historical experiment (1850-2014) from UKESM1-LL-0, CESM2-WACCM, MRI-ESM2-0, GISS-E2.1-G and GFDL-ESM4 and for ScenarioMIP SSP3-7.0 experiment (2015-2100) are represented with their intermodel standard deviation ($\pm 1 \sigma$, shaded areas). Observation based global tropospheric ozone burden estimate (from Table 6.3) is for 2010-2014. Tropospheric Ozone Assessment Report (TOAR) multi-model mean value (from Table 6.3) is for 2000 with a $\pm 1 \sigma$ error-bar. Atmospheric Chemistry and Climate Model Intercomparison Project (ACCMIP) multi-model means are for 1850, 1930, 1980 and 2000 time-slices with $\pm 1 \sigma$ error-bars. The troposphere is masked by the tropopause pressure calculated in each model using the WMO thermal tropopause definition. Further details on data sources and processing are available in the chapter data table (Table 6.SM.1).

[END FIGURE 6.4 HERE]

1 Since the mid-20th century, the CMIP6 model ensemble shows a higher global trend (Figure 6.4). Since mid-
2 1990s, the trends are better documented by observations (see also Section 2.2.5.3) and indicate spatial
3 heterogeneity. In particular, in-situ observations at remote surface sites and in the lower free troposphere
4 indicate positive trends that are far more common than negative trends, especially in the northern tropics and
5 across southern and eastern Asia (Figure 6.5). The CMIP6 ensemble and observations largely agree in the
6 magnitude of the global positive trend since 1997 (0.82 ± 0.13 Tg yr⁻¹ in the model ensemble; 0.70 ± 0.15 Tg
7 yr⁻¹ in ozonesonde dataset; 0.83 ± 0.85 Tg yr⁻¹ in the satellite ensemble) and qualitatively reproduce positive
8 trends in Southern Hemisphere (Griffiths et al., 2020). More analyses are needed for evaluation in other parts
9 of the world to assess the skill of the recent ensemble based on CMIP6 emissions.

10
11
12 **[START FIGURE 6.5 HERE]**

13
14 **Figure 6.5: Decadal tropospheric ozone trends since 1994.** Trends are shown at 28 remote and regionally
15 representative surface sites (Cooper et al., 2020) and in 11 regions of the lower free troposphere (650 hPa,
16 ca. 3.5 km) as measured by In-Service Aircraft for a Global Observing System (IAGOS) above Europe,
17 northeastern USA, southeastern USA, western North America, northeast China, South East Asia, southern
18 India, Persian Gulf, Malaysia/Indonesia, Gulf of Guinea and northern South America (Gaudel et al.,
19 2020). High elevation surface sites are >1500 m above sea level. All trends end with the most recently
20 available year but begin in 1995 or 1994. The sites and datasets are the same as those used in Figure 2.8,
21 further details on data sources and processing are available in the Chapter 2 data table (Table 2.SM.1).

22
23 **[END FIGURE 6.5 HERE]**

24
25
26 In summary, there is *high confidence* in the estimated present-day (about 2010) global tropospheric ozone
27 burden based on an ensemble of models and observational estimates (347 ± 28 Tg), but there is *medium*
28 *confidence* among the individual models for their estimates of the tropospheric ozone related budget terms.
29 Evidence from successive multi-model intercomparisons and the limited isotopic evidence agree on the
30 magnitude of the increase of the tropospheric ozone burden from 1850 to present-day in response to
31 anthropogenic changes in ozone precursor emissions corroborating AR5 findings. This increase is assessed
32 to be 109 ± 25 Tg (*medium confidence*). CMIP6 model ensemble shows a constant global increase since the
33 mid-20th century whose rate is consistent with that derived from observations since the mid-1990s.

34 35 36 6.3.2.2 Stratospheric ozone

37
38 Ninety percent of total-column ozone resides in the stratosphere. The chemical lifetime of ozone in the
39 stratosphere ranges from less than a day in the upper stratosphere to several months in the lower stratosphere
40 (Bekki and Lefevre, 2009). Global stratospheric ozone trends based on observations are assessed in Chapter
41 2 (Section 2.2.5.2). CMIP6 model ensemble shows that global total ozone column (TCO) has slightly
42 changed from 1850 to 1960 (Keeble et al., 2021). The rapid decline in the 1970s and 1980s due to
43 halogenated ODSs (as assessed in Chapter 2 Section 2.2.5.2 from observations) until the end of the 1990s
44 followed by a slight increase since then is captured by the models (Keeble et al., 2021). Overall the observed
45 climatology patterns and annual cycle amplitudes, are well represented in the CMIP6 ensemble mean.
46 CMIP6 ensemble overestimates the observed TCO values by up to 6% (10-20 DU) globally in the NH and
47 SH mid-latitudes, and in the tropics, but the trend in these regions is well captured between 1960 and 2014.
48 However, there is poor agreement between the individual CMIP6 models in the pre-industrial and throughout
49 the historical period, with model TCO values spread across a range of ~60 DU. The global stratospheric
50 ozone column decreased by 14.3 ± 8.7 DU from 1850 to 2014 (Keeble et al., 2021).

51
52 Model simulations attribute about half of the observed upper stratospheric ozone increase after 2000 to the
53 decline of ODS since the late 1990s while the other half of the ozone increase is attributed to the slowing of
54 gas-phase ozone destruction cycles due to cooling of the upper stratosphere by increasing GHGs (Aschmann
55 et al., 2014; Oberländer-Hayn et al., 2015).

1 In summary, global stratospheric ozone column has decreased from pre-industrial to present-day in response
2 to the ODS-induced ozone rapid decline in the 1970s and 1980s followed by slow, and still incomplete,
3 recovery. There is *medium confidence* that global stratospheric ozone column has changed by 14.3 ± 8.7 DU
4 between 1850 and 2014.

6.3.3 Precursor Gases

6.3.3.1 Nitrogen Oxides (NO_x)

11 The distribution of tropospheric NO_x is highly variable in space and time owing to its short lifetime coupled
12 with highly heterogeneous emission and sink patterns. NO_x undergoes chemical processing, including the
13 formation of nitric acid (HNO_3), nitrate (NO_3^-), organic nitrates (e.g. alkyl nitrate, peroxyacyl nitrate), and
14 atmospheric transport, and deposition. Despite challenges in retrieving quantitative information from satellite
15 observations (Duncan et al., 2014; Lin et al., 2015; Lorente et al., 2017; Silvern et al., 2018), improved
16 accuracy and resolution of satellite-derived tropospheric NO_2 columns over the past two decades have
17 advanced understanding of the global distribution, long-term trends, and source attribution of NO_x . Long-
18 term average tropospheric NO_2 column based on multiple satellites born instruments (Figure 6.6a) reveals
19 highest NO_2 levels over the most populated, urbanized, and industrialised regions of the world corresponding
20 to high NO_x emission source regions (Krotkov et al., 2016; Georgoulias et al., 2019). Enhanced but highly
21 variable NO_2 columns are also associated with biomass burning regions as well as areas influenced by
22 lightning activity (Miyazaki et al., 2014; Tanimoto et al., 2015).

23
24 Observational constraints derived from isotopic composition of atmospheric nitrate inferred from ice cores
25 provide evidence of increasing anthropogenic NO_x sources since pre-industrial times (Hastings et al., 2009;
26 Geng et al., 2014). Global NO_x emission trends in bottom-up inventories (Section 6.2.1) as well as model
27 simulations of nitrogen deposition (Lamarque et al., 2013b) are in qualitative agreement with these
28 observational constraints. CMIP6 ESMs exhibit stable NO_x burden until early 1900s and then a sharp
29 increase driven by a factor of three increase in emissions, however the magnitude of this increase remains
30 uncertain due to poor observational constraints on pre-industrial concentrations of NO_x (Griffiths et al.,
31 2020).

32
33 AR5 reported NO_2 decreases by 30-50% in Europe and North America and increases by more than a factor
34 of two in Asia over the 1996 to 2011 based on satellite observations (Hartmann et al., 2013). Extension of
35 this analysis covering time period up to 2015 reveals that NO_2 has continued to decline over the USA,
36 western Europe, and Japan (Schneider et al., 2015; Duncan et al., 2016; Krotkov et al., 2016) because of
37 effective fossil fuel NO_x emission controls (Section 6.2), although this rate of decline has slowed down post-
38 2011 (Jiang et al., 2018b). Satellite observations also reveal a 32% decline in NO_2 column over China after
39 peaking in 2011 (see Figure 6.6b) consistent with declining NO_x emissions (Section 6.2) due to the
40 implementation of emission control strategies (de Foy et al., 2016; Irie et al., 2016; Liu et al., 2016a). Over
41 South Asia, tropospheric NO_2 levels have grown rapidly with increases of 50% during 2005 to 2015 largely
42 driven by hotspot areas in India experiencing rapid expansion of the power sector (Duncan et al., 2016;
43 Krotkov et al., 2016). Further analysis indicates that many parts of India have also undergone a reversal in
44 NO_2 trends since 2011 that has been attributed to a combination of factors, including a slowdown in
45 economic growth, implementation of cleaner technologies, non-linear NO_x chemistry, and meteorological
46 variability (Georgoulias et al., 2019). Satellite data reveals spatially heterogeneous NO_2 trends over the
47 Middle East with an overall increase over 2005-2010 and a decrease over large parts of the region after
48 2011-2012. The reasons for trend reversal within individual areas are diverse, including warfare, imposed
49 sanctions, and air quality controls (Lelieveld et al., 2015a; Georgoulias et al., 2019). Satellite-derived
50 tropospheric NO_2 levels over Africa and Latin America do not show a clear trend; both increasing and
51 decreasing trends are observed over large agglomerations in these regions since the early 2000s (Schneider et
52 al., 2015; Duncan et al., 2016).

53
54 In summary, global tropospheric NO_x abundance has increased from 1850 to 2015 (*high confidence*).
55 Satellite observations of tropospheric NO_x indicate strong regional variations in trends over 2005-2015.

1 There is *high confidence* that NO₂ has declined over the USA and western Europe since the mid 1990s and
2 increased over China until 2011. NO₂ trends have reversed (declining) over China beginning in 2012 and
3 NO₂ has increased over South Asia by 50% since 2005 (*medium confidence*).
4
5

6 **[START FIGURE 6.6 HERE]**
7

8 **Figure 6.6: Long term climatological mean (a) and time evolution (b) of tropospheric nitrogen dioxide (NO₂)**
9 **vertical column density.** Values come from the merged GOME/SCIAMACHY/GOME-2 (TM4NO2A
10 version 2.3) dataset for the period 1996-2016 (Georgoulias et al., 2019). Time evolution of NO₂ column
11 shown in panel (b) is normalized to the fitted 1996 levels for the 10 regions shown as boxes in panel (a).
12 Further details on data sources and processing are available in the chapter data table (Table 6.SM.1).
13

14 **[END FIGURE 6.6 HERE]**
15

16 6.3.3.2 Carbon Monoxide (CO)

17 About half of the atmospheric CO burden is due to its direct emissions and the remainder is due to
18 atmospheric oxidation of methane and NMVOCs. Reaction with OH is the primary sink of CO with a
19 smaller contribution from dry deposition.
20
21

22 Since AR5, advances in satellite retrievals (e.g. Worden et al., 2013; Warner et al., 2014; Buchholz et al.,
23 2021a), ground based column observations (e.g. Zeng et al., 2012; Té et al., 2016), airborne platforms (e.g.
24 Cohen et al., 2018; Petetin et al., 2018), surface measurement networks (e.g., Andrews et al., 2014; Petron et
25 al., 2019; Prinn et al., 2018; Schultz et al., 2015) and assimilation products (e.g., Deeter et al., 2017;
26 Flemming et al., 2017; Zheng et al., 2019) have resulted in better characterization of the present day
27 atmospheric CO distribution. Typical annual mean surface CO concentrations range from ~120 ppb in the
28 Northern Hemisphere to ~40 ppb in the Southern Hemisphere (Petron et al., 2019). The sub-regional patterns
29 in CO reflect the distribution of emission sources. Seasonal hotspots are linked to areas of biomass burning
30 in tropical South America, equatorial Africa, Southeast Asia, and Australia. A study using data assimilation
31 techniques estimates a global mean CO burden of 356 ± 27 Tg over the 2002-2013 period (Gaubert et al.,
32 2017).
33

34 Global models generally capture the global spatial distribution of the observed CO concentrations but have
35 regional biases of up to 50% (e.g., Emmons et al., 2020; Horowitz et al., 2020). Despite updated emissions
36 datasets, the global multi-model and single model simulations persistently underestimate observed CO
37 concentrations at northern high- and mid-latitudes as well as in the Southern Hemisphere but with smaller
38 biases compared with that in the Northern Hemisphere (Monks et al., 2015b; Naik et al., 2013b; Stein et al.,
39 2014; Strode et al., 2015). Models are biased high in the tropics, particularly over highly polluted areas in
40 India and Eastern Asia (Strode et al., 2016; Yarragunta et al., 2017).
41

42 Estimates of global CO burden simulated by global models generally fall within the range of that derived
43 from data assimilation techniques, though the spread across the models is large (Myriokefalitakis et al.,
44 2016; Naik et al., 2013b; Stein et al., 2014; Zeng et al., 2015). There is a large diversity in model simulated
45 CO budget driven by uncertainties in CO sources and sinks, particularly those related to in situ production
46 from NMVOCs and loss due to reaction with OH (Stein et al., 2014; Zeng et al., 2015; Myriokefalitakis et
47 al., 2016). Global CO budget analysis from a multimodel ensemble for more recent years including results
48 from the CMIP6 model runs are not yet available.
49

50 Reconstructions of CO concentrations based on limited ice core samples in the Northern Hemisphere high
51 latitudes suggest CO mole fractions of about 145 ppb in the 1950s, which rose by 10-15 ppb in the mid-
52 1970s, and then declined by about 30 ppb to about 130 ppb by 2008 (Petrenko et al., 2013). The negative
53 trends since the 1990s are often attributed to emission regulations from road transportation in North America
54 and Europe. Due to limited observations prior to the satellite era, our knowledge on long-term global CO
55

1 trends are estimated using models. An increase of global CO burden of about 50% for the year 2000 relative
2 to 1850 is found in CMIP6 (Griffiths et al., 2020).

3
4 AR5 reported a global CO decline of about 1% yr⁻¹ based on satellite data from 2002-2010, but biases in
5 instruments rendered low confidence in this trend. AR5 also indicated a small CO decrease from in situ
6 networks but did not provide quantitative estimates. New analysis of CO trends performed since AR5 and
7 based on different observational platforms and assimilation products show a decline globally and over most
8 regions during the last one to two decades with varying amplitudes partly depending on the period of
9 analysis (Table 6.4). Inversion based analysis attribute the global CO decline during the past two decades to
10 decreases in anthropogenic and biomass burning CO emissions despite probable increase in atmospheric CO
11 chemical production (Gaubert et al., 2017; Jiang et al., 2017; Zheng et al., 2019). Furthermore, (Buchholz et
12 al., 2021) report a slowdown in global CO decline in 2010-2018 compared to 2002-2010, although the
13 magnitude and sign of this change in the trend varies regionally. Global models prescribed with emissions
14 inventories developed prior to the CMIP6 inventory capture the declining observed CO trends over North
15 America and Europe but not over East Asia (Strode et al., 2016). CMIP6 models driven by CMIP6 emissions
16 simulate a negative trend in global CO burden over the 1990-2020 period (Griffiths et al., 2020), however
17 the simulated trends have not yet been evaluated against observations.

18
19 In summary, our understanding of present day global CO distribution has increased since AR5 with newer
20 and improved observations and reanalysis. There is *high confidence* that global CO burden is declining since
21 2000. Evidence from observational CO reanalysis suggests this decline to be driven by reductions in
22 anthropogenic CO emissions, however this is yet to be corroborated by global ESM studies with the most
23 recent emission inventories.

24
25
26 **[START TABLE 6.4 HERE]**

27
28 **Table 6.4:** Summary of the global CO trends based on model estimates and observations.

Analysis period	Trends : Regions	Reference/Methodology
Global / Hemispheric		
2003 to 2015	-0.86 % yr ⁻¹	(Flemming et al., 2017) Model assimilating MOPITT
2002 to 2013	-1.4% yr ⁻¹	(Gaubert et al., 2017) Model assimilating MOPITT
2002-2018	-0.50±0.3% yr ⁻¹ : 60°N – 60°S (MOPITT) -0.56±0.3% yr ⁻¹ ; -0.61±0.2% yr ⁻¹ : 0°-60°N -0.35±0.3% yr ⁻¹ ; -0.33±0.3% yr ⁻¹ : 0°-60°S	(Buchholz et al., 2021) Satellite Observations MOPITT; AIRS
2000-2017	-0.32±0.05% yr ⁻¹	(Zheng et al., 2019) Satellite Observations MOPITT
2003-2014	~ 0.5 to -2.5 ppb yr ⁻¹ : Northern Hemisphere ~ 0 to -0.5 ppb yr ⁻¹ : Southern Hemisphere	(Flemming et al., 2017) NOAA Carbon Cycle Cooperative Global Air Sampling Network
2001-2013	-0.80 ppb yr ⁻¹ to -2.19 ppb yr ⁻¹ : Northern Hemisphere (Upper Troposphere / Tropopause Layer)	(Cohen et al., 2018) IAGOS Airborne
Pacific / Tropic		
2004-2013 (Spring Mean)	-2.9 ± 2.6 ppb yr ⁻¹ : Mauna Loa (19.54°N, 155.58°W)	(Gratz et al., 2015) Ground based

2004-2013 (Spring Mean)	-2.6 ± 1.8 ppb yr ⁻¹ : Sand Island Midway (28.21°N, 177.38°W)	(Gratz et al., 2015) Ground based
Europe		
1996-2006	-0.45±0.16 %yr ⁻¹ : Jungfrauoch (46.6°N 8.0°E) -1.00±0.24 %yr ⁻¹ : Zugspitze (47.4°N 11.0°E) -0.62±0.19 %yr ⁻¹ : Harestua (60.2°N 10.8°E) 0.61±0.16 %yr ⁻¹ : Kiruna (67.8°N 20.4°E)	(Angelbratt et al., 2011) Ground based
2001-2011 May-Sep	-3.1±0.30 ppb yr ⁻¹ : Pico Mt.Obs (38.47 °N 28.40 °W) -1.4±0.20 ppb yr ⁻¹ : Mace Head, Ireland	(Kumar et al., 2013) Ground based
2002-2018	-0.89 ±0.1 %yr ⁻¹ : Europe (45°-55°N 0°-15°E)	(Buchholz et al., 2021) Satellite Observations MOPITT
North America		
2001-2010	-2.5 ppb yr ⁻¹ : Thompson Farm (43.11 °N 70.95 °W) -2.3 ppb yr ⁻¹ : Mt. Washington (44.27 °N 71.30 °W) +2.8 ppb yr ⁻¹ : Castle Springs (43.75 °N 71.35 °W) -3.5 ppb yr ⁻¹ : Pack Monadnock (42.86 °N 71.88 °W) -2.8 ppb yr ⁻¹ : Whiteface Mountain (44.40 °N 73.90 °W) -4.3 ppb yr ⁻¹ : Pinnacle State Park (42.09 °N 77.21 °W)	(Zhou et al., 2017) Ground based
2004-2013 (Spring Mean)	-3.2 ± 2.9 ppb yr ⁻¹ : Mt. Bachelor Observatory	(Gratz et al., 2015) Ground based
2004-2012 (Spring Mean)	-2.8 ± 1.8 ppb yr ⁻¹ : Shemya Island (55.21°N, 162.72°W)	(Gratz et al., 2015) Ground based
2002-2018	-0.85 ± 0.1%yr ⁻¹ : Eastern USA (35°-40°N, -95° to -75°E)	(Buchholz et al., 2021) Satellite Observations MOPITT
Asia		
2005-2018	-0.46±0.14 % yr ⁻¹ : East Asia	(Zheng et al., 2018a) WDCGG Ground based
2005-2018	-0.41±0.09 % yr ⁻¹ : East Asia	(Zheng et al., 2018a) MOPITT
2002-2018	-1.18 ± 0.3 %yr ⁻¹ : (N.E. China 30°-40°, 110°-123°E) -0.28±0. : (N. India 20°-30°N 70°-95°E)	(Buchholz et al., 2021) Satellite Observations MOPITT

1
2 [END TABLE 6.4 HERE]
3
4

5 6.3.3.3 *Non-Methane Volatile Organic Compounds (NMVOCs)*
6

7 NMVOCs encompass thousands of compounds with lifetimes from hours to days to months and abundances
8 and chemical composition highly variable with respect to space, time. Although the biogenic source (Section
9 6.2.2) dominates the global NMVOC budget, anthropogenic activities are the main driver of long-term trends
10 in the abundance of many compounds.
11

12 Information on the global distribution of individual NMVOCs is scarce, except for the less reactive

1 compounds having lifetimes of several days to months. Based on measurements from polar firn air samples
2 and ground-based networks, AR5 reported that the abundances of the predominantly anthropogenic light
3 alkanes (C₂-C₅) increased until 1980 and declined afterwards. The decline was attributed to air quality
4 emission controls and to fugitive emission decreases following the collapse of the Soviet Union (Simpson et
5 al., 2012). Since AR5, scarce ground-based measurements have shown that the decline in C₂-C₃ alkanes
6 ended around 2008 and their abundances are since growing again, which is primarily attributed to increasing
7 North American emissions (see Section 6.2.1). Furthermore, since AR5 the evolution of ethane levels during
8 the past millennium was made accessible by analysis of ice core samples (Nicewonger et al., 2016). The
9 large observed interpolar ratio of ethane in preindustrial times (3.9) corroborates a large geologic source of
10 ethane previously put forward by (Etiope and Ciccioli, 2009), and narrows down its likely global magnitude
11 (Nicewonger et al., 2018) (*low to medium confidence*). The incorporation of geologic emissions in CCMs is
12 not yet systematic though a one-model study has shown improved agreement of the results with observations
13 (Dalsøren et al., 2018).

14
15 Formaldehyde is a short-lived high-yield product of NMVOC oxidation and formaldehyde column data from
16 satellite instruments can therefore inform on trends in anthropogenic NMVOC abundances over very
17 industrialized region. AR5 reported significant positive trends in HCHO between 1997 and 2009 over north-
18 eastern China (4% yr⁻¹) and negative trends over north-eastern U.S. cities. Since AR5, there is *robust*
19 *evidence* and *high agreement* of an upward trend of HCHO over eastern China, though large regional
20 disparities exist in the trends (De Smedt et al., 2015; Shen et al., 2019) with possible negligible or decreasing
21 trend over Beijing and the Pearl River Delta. In other world regions, in particular North America, there is
22 *limited to medium evidence* for significant changes in the HCHO columns, except in regions where the trend
23 is particularly strong, e.g. the Houston area (-2.2% yr⁻¹ over 2005-2014) and the Alberta oil sands (+3.8% yr⁻¹)
24 (Zhu et al., 2017). Over the north-eastern U.S., even the sign of the trend differs between studies (De
25 Smedt et al., 2015; Zhu et al., 2017) for reasons that are unclear.

26
27 In summary, after a decline between 1980 and 2008, abundances of light NMVOCs have increased again
28 over the Northern Hemisphere due to extraction of oil and gas in North America (*high confidence*). Trends in
29 satellite HCHO observations, used as a proxy of anthropogenic NMVOC over industrialised areas, show a
30 significant positive trend over Eastern China (*high confidence*) but also indicate large regional disparities in
31 the magnitude of the trends over China and even in their signs over North America.

32 33 34 6.3.3.4 Ammonia (NH₃)

35
36 Ammonia is the most abundant alkaline gas in the atmosphere. Its present-day source is dominated by
37 livestock and crop production (section 6.2). Ammonia reacts with nitric acid and sulphuric acid to produce
38 ammonium sulphate and ammonium nitrate, which contribute to the aerosol burden (see section 6.3.5.2),
39 promotes aerosol nucleation by stabilizing sulphuric acid clusters (Kirkby et al., 2011) and contributes to
40 nitrogen deposition (see section 6.4.4) (Sheppard et al., 2011; Flechard et al., 2020). Trends in NH₃ were not
41 assessed in AR5.

42
43 Considerable expansion of satellite (Clarisse et al., 2009; Sheppard and Cady-Pereira, 2015; Warner et al.,
44 2016) and ground-based observations (Miller et al., 2014; Li et al., 2016; Pan et al., 2018) has improved our
45 understanding of the spatial distribution and seasonal to inter-annual variability of ammonia, and advanced
46 its representations in models (e.g., Zhu et al., 2015). Regionally, peak NH₃ concentrations are observed over
47 large agricultural (e.g., Northern India, US Midwest and Central Valley) and biomass burning regions, in
48 good qualitative agreement with emission inventories (Van Damme et al., 2015, Van Damme et al., 2018).
49 However, several large agricultural and industrial hotspots have been found to be missing or greatly
50 underestimated in emission inventories (Van Damme et al., 2018). NH₃ exhibits a strong vertical gradient,
51 with a maximum in the boundary layer (Schiferl et al., 2016) and can be transported into the upper
52 troposphere and lower stratosphere (UTLS), particularly in the Asian Monsoon region, as indicated by
53 observations (Froyd et al., 2009; Höpfner et al., 2016, 2019) and theoretical considerations (Ge et al., 2018).
54 There is a large range in the present-day NH₃ burden (from 0.04 to 0.7 TgN) simulated by CCMs,
55 highlighting deficiencies in the process-level representation of NH₃ in current global models (Bian et al.,

2017). The underestimate of surface NH₃ concentrations (Bian et al., 2017) further highlights such deficiencies and the limitations in comparing site-specific observations with relatively coarse-resolution models.

Observations show that NH₃ concentration has been increasing in recent decades in the USA (Butler et al., 2016; Warner et al., 2016; Yu et al., 2018), western Europe (van Zanten et al., 2017; Warner et al., 2017, Wichink Kruit et al., 2017; Tang et al., 2018b), and China (Warner et al., 2017; Liu et al., 2018). This trend has been attributed to a combination of increasing ammonia emissions (Sutton et al., 2013; Fowler et al., 2015) and decreases in the chemical reaction of NH₃ with nitric and sulphuric acids associated with reductions in SO₂ and NO_x emissions whose rate depends on the region (Warner et al., 2017; Yao and Zhang, 2019). Over longer time scales, CCMs simulate an increase of the NH₃ burden by a factor of 2 to 7 since preindustrial conditions (Xu and Penner, 2012; Hauglustaine et al., 2014).

In summary, progress has been made in the understanding of the spatio-temporal distribution of ammonia, though representation of NH₃ remains rather unsatisfactory due to process-level uncertainties. Evidence from observations and models suggests that ammonia concentrations have been increasing over the recent decades due to emissions and chemistry. There is *high confidence* that the global NH₃ burden has increased considerably from preindustrial to present-day although the magnitude of the increase remains uncertain.

6.3.3.5 Sulphur Dioxide (SO₂)

AR5 did not assess trends in SO₂ concentrations. Trends in SO₂ abundances are consistent with the overall anthropogenic emission changes as presented in Section 6.2 and Figure 6.18. Long-term surface-based in situ observations in North America and Europe show reductions of more than 80% since the measurements began around 1980 (Table 6.5). Europe had the largest reductions in the first part of the period while the highest reduction came later in North America. Observed trends are qualitatively reproduced by global and regional models over North America and Europe over the period 1990-2015 for which emission changes are well quantified (Table 6.5) (Aas et al., 2019).

In situ observations over other parts of the world are scattered. However, the limited in-situ observations in East Asia indicate an increase in atmospheric SO₂ up to around 2005 and then a decline (Aas et al., 2019). This is confirmed by satellite observations (Krotkov et al., 2016), which further reveal a rapid decline in SO₂ since 2012 to 2013 (Krotkov et al., 2016; Zheng et al., 2018b). In India, on the other hand, the SO₂ levels have doubled over 2005 to 2015 (Krotkov et al., 2016).

In summary, surface and satellite observations indicate strong regional variations in trends of atmospheric SO₂ abundance. The SO₂ concentrations in North America and Europe have declined over 1980 to 2015 with slightly stronger reductions in North America (70 ± 20%) than over Europe (58 ± 32%) over 2000-2015, though Europe had larger reductions than the US in the prior decade (1990-2000). In Asia, the SO₂ trends are more scattered, though there is *medium confidence* that there was a strong increase up to around 2005, followed by a steep decline in China, while over India, the concentrations are increasing steadily.

[START TABLE 6.5 HERE]

Table 6.5: Summary of changes or trends in atmospheric abundance of sulphur dioxide (SO₂) and sulphate (SO₄²⁻) aerosols based on in situ and satellite observations.

Analysis period	Trends in SO ₂	Trends in particulate SO ₄ ²⁻	Reference
Global Models/Assimilated models			

1990-2000	-8.54 ± 1.40 % yr ⁻¹ (EU, 43 sites) -2.63 ± 0.30 % yr ⁻¹ (NA, 53 sites)	-5.23 ± 1.17% yr ⁻¹ (EU, 41 sites) -1.94 ± 0.43% yr ⁻¹ (NA 101 sites)	(Aas et al., 2019)
2000-2015	-0.41 ± 0.92 % yr ⁻¹ (EA, 19 sites) -4.86 ± 1.31 % yr ⁻¹ (EU, 47 sites) -4.40 ± 0.93 % yr ⁻¹ (NA, 77 sites)	0.02 ± 0.91% yr ⁻¹ (EA, 13 sites) -3.26 ± 0.85% yr ⁻¹ (EU, 36 sites) -3.18 ± 0.66% yr ⁻¹ (NA, 218 sites)	(Aas et al., 2019)
Ground based in situ observations			
1980-1990	-5.03 ± 2.04 % yr ⁻¹ (EU, 20 sites) -2.5 % yr ⁻¹ (US)	-2.56 ± 3.10% yr ⁻¹ (EU, 16 sites) -1.80 ± 4.09% yr ⁻¹ (US SO ₄ ²⁻ in precipitation, 78 sites)	(Aas et al., 2019) US EPA^a
1990-2000	-7.56 ± 1.81 % yr ⁻¹ (EU, 43 sites) -3.27 ± 1.69 % yr ⁻¹ (NA, 53 sites)	-5.16 ± 2.11% yr ⁻¹ (EU, 41 sites) -2.08 ± 1.44% yr ⁻¹ (NA, 101 sites)	(Aas et al., 2019)
2000-2015	-0.14 ± 5.32 % yr ⁻¹ (EA, 19 sites) -3.89 ± 2.16 % yr ⁻¹ (EU, 47 sites) -4.69 ± 1.35 % yr ⁻¹ (NA, 77 sites)	2.68 ± 9.41% yr ⁻¹ (EA, 13 sites) -2.67 ± 2.03% yr ⁻¹ (EU, 36 sites) -3.15 ± 1.30% yr ⁻¹ (NA, 218 sites)	(Aas et al., 2019)
Change based on satellite observations			
2005-2015	ca -80% (Eastern US)		(Krotkov et al., 2016)
2005-2015	ca -60% (Eastern EU)		(Krotkov et al., 2016)
2005-2015	200 ± 50 % (India)		(Krotkov et al., 2016)
2005 (and 2012) - 2015	ca -50 % (The North China Plain)		(Krotkov et al., 2016)

^a <https://www.epa.gov/air-trends/sulfur-dioxide-trends>

[END TABLE 6.5 HERE]

6.3.4 Short-lived Halogenated Species

The halogenated species are emitted in the atmosphere in the form of the synthetically produced chlorofluorocarbons (CFCs), halons, hydrochlorofluorocarbons (HCFCs), hydrofluorocarbons (HFCs) and others. Their historical global abundances are provided in Annex III and discussed in Chapter 2 (Section 2.2.4, Table 2.3). In summary, for the period 2011-2019, the abundance of total chlorine from HCFCs has continued to increase in the atmosphere with decreased growth rates, total tropospheric bromine from halons and methyl bromide continued to decrease while abundances of most currently measured HFCs increased significantly, consistent with expectations based on the ongoing transition away from the use of ODSs. Here, emphasis is given on the very short-lived halogenated species (VSLs). The trends for these species were not discussed in IPCC AR5.

VSLs are halogenated substances with atmospheric lifetimes less than 0.5 year. While longer-lived ODSs account for most of the present day stratospheric halogen loading, there is *robust evidence* that VSLs contribute to stratospheric bromine and chlorine (Carpenter et al., 2014; Hossaini et al., 2015; Leedham Elvidge et al., 2015b) thus also contributing to stratospheric ozone depletion.

Of the atmospheric VSLs, brominated and iodinated species are predominantly of oceanic origin, while chlorinated species have significant additional anthropogenic sources (Carpenter et al., 2014; Hossaini et al., 2015). Global mean chlorine from the VSLs has increased in the troposphere from about 91 ppt in 2012 to about 110 ppt in 2016 (Engel et al., 2018). This increase is mostly due to dichloromethane (CH₂Cl₂), a species that has predominantly anthropogenic sources reflected by 3-times higher concentrations in the Northern Hemisphere than in the Southern Hemisphere (Hossaini et al., 2017). The upward dichloromethane

1 trend is corroborated by upper tropospheric aircraft data over the period 1998-2014 (Leedham Elvidge et al.,
2 2015a; Oram et al., 2017). The observations from the surface networks show that the abundance of
3 dichloromethane continued to increase until 2019 (Annex III), although the accuracy of global abundance of
4 VSLs is limited by the limited coverage by networks. No long-term change of the bromine containing
5 VSLs have been observed (Engel et al., 2018).

6.3.5 Aerosols

10 This section assesses trends in atmospheric distribution of aerosols and improvements in relevant physical
11 and chemical processes. The observed large-scale temporal evolution of aerosols is assessed in Section 2.2.6.
12 Since AR5, long term measurements of aerosol mass concentrations from regional global surface networks
13 have continued to expand and provide information on the distribution and trends in aerosols (Figure 6.7).
14 There is large spatial variability in aerosol mass concentration, expressed as PM_{2.5}, dominant aerosol type
15 and aerosol composition, consistent with the findings in AR5.

18 [START FIGURE 6.7 HERE]

20 **Figure 6.7: Distribution of PM_{2.5} composition mass concentration (in $\mu\text{g m}^{-3}$) for the major PM_{2.5} aerosol**
21 **components.** Those aerosol components are sulphate, nitrate, ammonium, sodium, chloride, organic
22 carbon, and elemental carbon. The central world map depicts the intermediate level regional breakdown
23 of observations (10 regions) following the IPCC Sixth Assessment Report Working Group III (AR6
24 WGIII). Monthly averaged PM_{2.5} aerosol component measurements are from: (i) the Environmental
25 Protection Agency (EPA) network which include 211 monitor sites primarily in urban areas of North
26 America during 2000-2018 (Solomon et al., 2014) (ii) the Interagency Monitoring of Protected Visual
27 Environments (IMPROVE) network during 2000-2018 over 198 monitoring sites representative of the
28 regional haze conditions over North America, (iii) the European Monitoring and Evaluation Programme
29 (EMEP) network over 70 monitoring in Europe and (eastern) Eurasia during 2000-2018, (iv) the Acid
30 Deposition Monitoring Network in East Asia (EANET) network with 39 (18 remote, 10 rural, 11 urban)
31 sites in Eurasia, Eastern Asia, South-East Asia and Developing Pacific, and Asia-Pacific Developed
32 during 2001-2017, (v) the global Surface Particulate Matter Network (SPARTAN) during 2013-2019
33 with sites primarily in highly populated regions around the world (i.e, North America, Latin America and
34 Caribbean, Africa, Middle East, Southern Asia, Eastern Asia, South-Eastern Asia and Developing
35 Pacific) (Snider et al., 2015, 2016), and (vii) individual observational field campaign averages over Latin
36 America and Caribbean, Africa, Europe, Eastern Asia, and Asia-Pacific Developed (Celis et al., 2004;
37 Feng et al., 2006; Mariani and de Mello, 2007; Molina et al., 2007, 2010; Bourotte et al., 2007; Fuzzi et
38 al., 2007; Mkoma, 2008; Favez et al., 2008; Aggarwal and Kawamura, 2009; Mkoma et al., 2009; Li et
39 al., 2010; Martin et al., 2010; Radhi et al., 2010; Weinstein et al., 2010; de Souza et al., 2010; Batmunkh
40 et al., 2011; Pathak et al., 2011; Gioda et al., 2011; Zhang et al., 2012a; Zhao et al., 2013; Cho and Park,
41 2013; Wang et al., 2019; Kuzu et al., 2020). Further details on data sources and processing are available
42 in the chapter data table (Table 6.SM.1).

44 [END FIGURE 6.7 HERE]

47 Remote sensing instruments provide a larger-scale view of aerosol distributions and trends than ground-
48 based monitoring networks by retrieving the Aerosol optical depth (AOD), which is indirectly related to
49 aerosol mass concentrations. AOD is the column-integrated aerosol mass extinction at a given wavelength,
50 and is therefore relevant to the estimation of the radiative forcing of aerosol-radiation interactions (Section
51 7.3.3.1). Models participating in Phase III of the AeroCom intercomparison project were found to
52 underestimate present-day AOD by about 20% (Gliß et al., 2020), although different remote sensing
53 estimates obtain different estimates of global mean AOD. Gliß et al. (2020) also highlight the considerable
54 diversity in the simulated contribution of various aerosol types to total AOD. However, models simulate
55 regional trends in AODs that agree well, when expressed as percentage change, with ground- (Gliß et al.,
56 2020; Mortier et al., 2020) and satellite-based (Cherian and Quaas, 2020a; Gliß et al., 2020) observations.
57 AOD trends simulated by CMIP6 models are more consistent with satellite-derived trends than CMIP5

1 models for several subregions, thanks to improved emission estimates (Cherian and Quaas, 2020b).

2
3 All CMIP6 models simulate a positive trend in global mean AOD since 1850, with a strong increase after the
4 1950s coinciding with the massive increase in anthropogenic SO₂ emissions (Figure 6.8). Global mean AOD
5 increases have slowed since 1980, or even reversed in some models, as a result of a compensation between
6 SO₂ emission decreases over the United States and Europe in response to air quality controls since the mid-
7 1980s and increases over Asia. Since the mid-2000s, global mean AOD stabilized driven by soaring
8 emissions in South Asia, and declining emissions in East Asia (section 6.2.1). Trends post ~2010 are difficult
9 to assess from CMIP6 models since the historical simulations end at 2014. Nevertheless, the strong decline
10 in anthropogenic SO₂ emissions over East Asia since 2011 is underestimated in the CMIP6 emission
11 database (Hoesly et al., 2018) indicating that the observed AOD change over East Asia may not be captured
12 accurately by CMIP6 models (Wang et al., 2021). While all CMIP6 models simulate increase of AOD
13 between 1850 and 2014 there is strong inter-model diversity in the simulated AOD change since 1850
14 ranging from 0.01 (15%) to 0.08 (53%) in 2014. Some models therefore lie outside the 68% confidence
15 interval of 0.02 (15%) to 0.04 (or 30%) for global AOD change in 2005-2015 compared to 1850 estimated
16 by (Bellouin et al., 2020) based on observational and model (excluding CMIP6) lines of evidence. In
17 addition to the horizontal distribution of aerosols documented by AOD, their number size distribution,
18 vertical distribution, optical properties, hygroscopicity, ability to act as CCN, chemical composition, mixing
19 state and morphology are key elements to assess their climate effect (see Section 6.4.).

20
21
22 **[START FIGURE 6.8 HERE]**

23
24 **Figure 6.8: Time evolution of changes in global mean aerosol optical depth (AOD) at 550 nm.** Year of reference
25 is 1850. Data shown from individual Coupled Model Intercomparison Project Phase 6 (CMIP6) historical
26 simulations. Each time series corresponds to the ensemble mean of realizations done by each model.
27 Simulation results from years including major volcanic eruptions, e.g. Novarupta (1912) and Pinatubo
28 (1991), are excluded from the analysis for models encompassing the contribution of stratospheric
29 volcanic aerosols to total AOD. Further details on data sources and processing are available in the chapter
30 data table (Table 6.SM.1).

31
32 **[END FIGURE 6.8 HERE]**

33 34 35 6.3.5.1 Sulphate (SO₄²⁻)

36
37 Sulphate aerosols (or sulphate containing aerosols) are emitted directly or formed in the atmosphere by gas
38 and aqueous phase oxidation of precursor sulphur gases, including SO₂, DMS, and carbonyl sulphide (OCS),
39 emitted from anthropogenic and natural sources (Section 6.2). Sulphate aerosols influence climate forcing
40 directly by either scattering solar radiation or absorbing longwave radiation, and indirectly by influencing
41 cloud micro- and macrophysical properties and precipitation (Boucher et al., 2013; Myhre et al., 2013b).
42 Additionally, sulphate aerosols and sulphate deposition have a large impact on air quality and ecosystems
43 (Reis et al., 2012). The majority of sulphate particles are formed in the troposphere, however, SO₂ and other
44 longer-lived natural precursors, such as OCS, transported into the stratosphere contribute to the background
45 stratospheric aerosol layer (Kremser et al., 2016). SO₂ emissions from volcanic eruptions are a significant
46 source of stratospheric sulphate loading (see Chapter 2 for reconstruction of stratospheric aerosol optical
47 depth and Chapter 7 for radiative forcing of volcanic aerosols). Furthermore, studies suggest contributions
48 from anthropogenic SO₂ emissions transported into the stratosphere with a consequent impact on radiative
49 forcing (Myhre et al., 2004; Yu et al., 2016). However there is significant uncertainty in the relative
50 importance of this stratospheric sulphate source (Kremser et al., 2016).

51
52 Process understanding of sulphate production pathways from SO₂ emissions has seen some progress since
53 AR5. More specifically, many global climate models now have a more complete description of chemical
54 reactions such that oxidant levels (including ozone) are better described, include a pH-dependence of SO₂
55 oxidation (e.g., Bauer et al., 2020; Kirkevåg et al., 2018), and implement explicit descriptions of ammonium
56 and nitrate aerosol components, which may influence the partitioning of sulphate (Bian et al., 2017; Lund et

1 al., 2018). The pH influences the heterogeneous chemistry as well as the physical properties of the aerosols,
2 and this topic has been a subject of growing interest since AR5 (Cheng et al., 2016; Freedman et al., 2019;
3 Nenes et al., 2020). Increases in cloudwater pH have been shown to significantly increase the radiative
4 forcing of sulphates (Turnock et al., 2019).

5
6 Sulphate is removed from the atmosphere by dry deposition and wet scavenging, and these processes depend
7 on the characteristics of the Earth's surface, and the intensity, frequency and amount of precipitation
8 (Boucher et al., 2013). Even though there have been some improvements since AR5, model representation of
9 wet scavenging and related cloud processes, and atmospheric transport remains a key source of uncertainty
10 in the simulated aerosol distribution and lifetime with further consequences for the sulphate forcing estimates
11 (Kristiansen et al., 2016; Lund et al., 2018a). There are also still relatively large uncertainties in the emission
12 height used in models affecting the simulated aerosol distribution (Yang et al., 2019).

13
14 Based on long-term surface-based in situ observations, AR5 reported strong decline in sulphate aerosols in
15 Europe and the USA over 1990 to 2009, with the largest decreases occurring before 2000 in Europe and post
16 2000 in the USA. Since AR5, atmospheric measurements in conjunction with model results have provided
17 insights into the spatial and temporal distribution of sulphate and sulphur deposition (Vet et al., 2014; Tan et
18 al., 2018; Aas et al., 2019). The *in situ* observations in North America and Europe reveal substantial
19 reduction since the measurements started around 1980, though the trends have not been linear through this
20 period (Table 6.5). Several regional studies agree with these trend estimates for Europe (Banzhaf et al.,
21 2015b; Theobald et al., 2019) and North America (Sickles II and Shadwick, 2015; Paulot et al., 2016a).
22 Further, the concentrations of primary emitted SO₂ (Section 6.3.3.5) show greater decreases than secondary
23 sulphate aerosols over these regions due to a combination of higher oxidation rate (hence more SO₂
24 converted to SO₄²⁻) and increased dry deposition rate of SO₂ (Fowler et al., 2009; Banzhaf et al., 2015). In
25 situ observations over other parts of the world are scattered (see Figure 6.7), and the lack of observations
26 makes it too uncertain to quantify regional representative trends (Hammer et al., 2018). However, limited in-
27 situ observations in East Asia indicate an increase in atmospheric sulphate up to around 2005 and then a
28 decline (Aas et al., 2019) which is confirmed by satellite observations of SO₂ (Section 6.3.3.5). In India, on
29 the other hand satellite observations indicate a rapid increase in the SO₂ levels (Krotkov et al., 2016), and
30 long-term measurements of sulphate in precipitation in India further provide evidence of an increasing trend
31 from 1980 to 2010 (Bhaskar and Rao, 2017; Aas et al., 2019). Further improvements in global trend
32 assessments are expected with new integrated reanalysis products from the Earth-system data assimilation
33 projects (Randles et al., 2017; Inness et al., 2019).

34
35 Indirect evidence of decadal trends in the atmospheric loading of sulphur are provided by Alpine ice cores,
36 mainly influenced by European sources (Engardt et al., 2017), and ice cores from Svalbard (Samyn et al.,
37 2012) and Greenland (Patris et al., 2002; Iizuka et al., 2018) influenced by sources in Europe and North
38 America. These show similar patterns with a weak increase from the end of the 19th century up to around
39 1950 followed by a steep increase up to around 1980, and then a significant decrease over the next two
40 decades. This general trend is consistent with the emissions of SO₂ in North America and Europe (Hoesly et
41 al., 2018; Figures 6.18; 6.19).

42
43 Global and regional models qualitatively reproduce observed trends over North America and Europe for the
44 period 1990-2015 for which emission changes are generally well quantified (Aas et al., 2019; Mortier et al.,
45 2020), building confidence in the relationship between emissions, concentration, deposition and radiative
46 forcing derived from these models. Though, the models seem to systematically underestimate sulphate (Bian
47 et al., 2017; Lund et al., 2018a) and AOD (Lund et al., 2018a; Gliß et al., 2020), and there are quite large
48 differences in the models' distribution of the concentration fields of sulphate driven by differences in the
49 representation of photochemical production and sinks of aerosols. One global model study also highlighted
50 biases in simulated sulphate trends over the 2001-2015 period over eastern China due to uncertainties in the
51 CEDS anthropogenic SO₂ emissions trends (Paulot et al., 2018).

52
53 In summary, there is *high confidence* that the global tropospheric sulphate burden increased from 1850 to
54 around 2005, but there are large regional differences in the magnitude. Sulphate aerosol concentrations in
55 North America and Europe have declined over 1980 to 2015 with slightly stronger reductions in North

1 America (47%) than over Europe (40%) over 2000-2015, though Europe had larger reductions in the prior
2 decade (1990-2000), 52% and 21% respectively for Europe and North America. In Asia, the trends are more
3 scattered, though there is *medium confidence* that there was a strong increase up to around 2005, followed by
4 a steep decline in China, while over India, the concentrations are increasing steadily.

6.3.5.2 Ammonium (NH_4^+), and Nitrate Aerosols (NO_3^-)

9 Ammonium sulphate and ammonium nitrate aerosols are formed when NH_3 reacts with nitric acid (HNO_3)
10 and sulphuric acid (H_2SO_4) produced in the atmosphere by the oxidation of NO_x and SO_2 , respectively.
11 Ammonium nitrate is formed only after H_2SO_4 is fully neutralized. NH_4^+ and NO_3^- aerosols produced via
12 these gas-to-particle reactions are a major fraction of fine-mode particles (with diameter $< 1\mu\text{m}$) affecting air
13 quality and climate. Coarse-mode nitrate, formed by the heterogeneous reaction of nitric acid with dust and
14 sea salt, dominates the overall global nitrate burden, but has little radiative effect (Hauglustaine et al., 2014;
15 Bian et al., 2017). Trends in ammonium (NH_4^+) and nitrate (NO_3^-) were not assessed in AR5.

17 Global model present-day estimates of the global NH_4^+ burden range from 0.1 to 0.6 TgN (Bian et al., 2017).
18 Models generally simulate surface NH_4^+ concentrations better than surface NH_3 concentrations (Bian et al.,
19 2017), which reflects its thermodynamic control by SO_4^{2-} rather than NH_3 (Shi et al., 2017). The concomitant
20 increases of NH_3 , SO_2 , and NO_x emissions (see Section 6.2) have led to a factor of 3 to 9 increase in the
21 simulated NH_4^+ burden from 1850 to 2000 (Hauglustaine et al., 2014; Lund et al., 2018a), driven primarily
22 by ammonium sulphate (70-90%). The increases in the NH_3 and NH_4^+ burdens are indirectly supported by
23 the observed increase of NH_4^+ concentration in ice cores in mid to high latitudes (Kang et al., 2002; Kekonen
24 et al., 2005; Lamarque et al., 2013a; Iizuka et al., 2018).

26 Ammonium nitrate is semi-volatile, which results in complex spatial and temporal patterns in its
27 concentrations (Putaud et al., 2010; Hand et al., 2012a; Zhang et al., 2012b) reflecting variations in its
28 precursors, NH_3 and HNO_3 , as well as SO_4^{2-} , non-volatile cations, temperature and relative humidity (Nenes
29 et al., 2020). High relative humidity and low temperature as well as elevated fine particulate matter loading
30 (Huang et al., 2014; Petit et al., 2015; Li et al., 2016; Sandrini et al., 2016) favour nitrate production.
31 Measurements reveal high contribution of NO_3^- to surface $\text{PM}_{2.5}$ ($>30\%$) in regions with elevated regional
32 NO_x and NH_3 emissions, such as the Paris area (Beekmann et al., 2015; Zhang et al., 2019), northern Italy
33 (Masiol et al., 2015; Ricciardelli et al., 2017), Salt Lake City (Kuprov et al., 2014; Franchin et al., 2018), the
34 North China Plains (Guo et al., 2014; Chen et al., 2016), and New Delhi (Pant et al., 2015). Recent
35 observations also show that ammonium nitrate contributes to the Asian tropopause aerosol layer (Vernier et
36 al., 2018; Höpfner et al., 2019). Model diversity in simulating present-day global fine-mode NO_3^- burden is
37 large with two multimodel intercomparison studies reporting estimates in the range of 0.14-1.88 Tg and
38 0.08-0.93 Tg, respectively (Bian et al., 2017; Gliß et al., 2021). Models differ in their estimates of the global
39 tropospheric nitrate burden by up to a factor of 13 with differences remain nearly the same across CMIP5
40 and CMIP6 generation of models (Bian et al., 2017; Gliß et al., 2021). While regional patterns in the
41 concentration of fine-mode NO_3^- are qualitatively captured by models, the simulation of fine-mode NO_3^- is
42 generally worse than that of NH_4^+ or SO_4^{2-} (Bian et al., 2017). This can be partly attributed to the semi-
43 volatile nature of ammonium nitrate and biases in the simulation of its precursors (Heald et al., 2014; Paulot
44 et al., 2016), including the sub-grid scale heterogeneity in NO_x and NH_3 emissions (Zakoura and Pandis,
45 2018).

47 Models indicate that the burden of fine-mode NO_3^- has increased by a factor of 2-5 from 1850 to 2000 (Xu
48 and Penner, 2012; Hauglustaine et al., 2014; Lund et al., 2018a), an increase that has accelerated between
49 2001 and 2015 (Lund et al., 2018a; Paulot et al., 2018). The sensitivity of NO_3^- to changes in NH_3 , SO_4^{2-} , and
50 HNO_3 is determined primarily by aerosol pH, temperature, and aerosol liquid water (Guo et al., 2016, 2018;
51 Weber et al., 2016; Nenes et al., 2020). In regions, where aerosol pH is high, changes in NO_3^- follow changes
52 in NO_x emissions, consistent with the observed increase of ammonium nitrate in Northern China from 2000
53 to 2015 (Wen et al., 2018) and its decrease in the US Central Valley (Pusede et al., 2016). In contrast, the
54 decrease in SO_2 emissions in the US Southeast has caused little change in NO_3^- from 1998 to 2014 as nitric
55 acid largely remains in the gas phase due to highly acidic aerosols (Weber et al., 2016; Guo et al., 2018).

In summary, there is *high confidence* that the NH_4^+ and NO_3^- burdens have increased from preindustrial to present-day, although the magnitude of the increase is uncertain especially for NO_3^- . The sensitivity of NH_4^+ and NO_3^- to changes in NH_3 , H_2SO_4 , and HNO_3 is well understood theoretically. However, it remains challenging to represent in models in part because of uncertainties in the simulation of aerosol pH and only a minority of ESMs consider nitrate aerosols in CMIP6.

6.3.5.3 Carbonaceous Aerosols

Carbonaceous aerosols are black carbon (BC)³, which is soot made almost purely of carbon, and organic aerosols⁴ (OA), which also contain hydrogen and oxygen and can be of both primary (POA) or secondary (SOA) origin. BC and a fraction of OA called Brown Carbon (BrC) absorb solar radiation. The various components of carbonaceous aerosols have different optical properties, so the knowledge of their partition, mixing, coating and ageing is essential to assess their climate effect (see Section 7.3.3.1.2).

Carbonaceous aerosols receive attention in the scientific and policy arena due to their radiative forcing, and their sizeable contribution to PM in an air quality context (Rogelj et al., 2014b; Harmsen et al., 2015; Shindell et al., 2016; Haines et al., 2017a; Myhre et al., 2017). BC exerts a positive ERF, but the ERF of carbonaceous aerosol as a whole is negative (Bond et al., 2013; Thornhill et al., 2021b). On average, carbonaceous aerosols accounts for 50 to 70% of PM with diameter lower than 1 μm in polluted and pristine areas (Zhang et al., 2007; Carslaw et al., 2010; Andreae et al., 2015; Monteiro dos Santos et al., 2016; Chen et al., 2017).

An extensive review on BC (Bond et al., 2013) discussed limitations in inferring its atmospheric abundance and highlighted inconsistencies between different terminology and related measurement techniques (Petzold et al., 2013; Sharma et al., 2017). Due to a lack of global observations, AR5 only reported declining total carbonaceous aerosol trends from USA and declining BC trend from the Arctic based on data available up to 2008. Since AR5, the number of observation sites has grown worldwide (see also Figure 6.7) but datasets suitable for global trend analyses remain limited (Reddington et al., 2017; Laj et al., 2020). Locally, studies based on observations from rural and background sites have reported decreasing surface carbonaceous aerosol trends in the Arctic, Europe, USA, Japan and India (Table 6.6). Increases in carbonaceous aerosol concentrations in some rural sites of the Western USA have been associated with wildfires (Hand et al., 2013; Malm et al., 2017). Long-term OA observations are scarce, so their trends outside of the USA are difficult to assess. Ice-core analysis have provided insight into carbonaceous aerosol trends predating the satellite and observation era over the Northern Hemisphere (Section 2.2.6, Figure 2.9b).

[START TABLE 6.6 HERE]

Table 6.6: Summary of the regional carbonaceous aerosol trends at background observation sites.

Species	Analysis period	Change/Trends	References
BC	1990-2009	Arctic Sites (Alert, Barrow, Ny-Alesund) -2% yr ⁻¹	(Sharma et al., 2013a)
	1970-2010	Finland (Kevo remote site) -1.8% yr ⁻¹	(Dutkiewicz et al., 2014)
	2005-2014	Germany (rural site) -2% yr ⁻¹	(Kutzner et al., 2018)

³ The terms EC and BC are operationally defined on the basis of the methodology used for their quantification, i.e., thermal refractivity and light absorption, respectively, and often used interchangeably.

⁴ The carbon mass fraction of OA is termed as OC, conversion factor between 1.4 to 1.6 from OC to OA are typically assumed

	2009-2016	United Kingdom (Harwell rural site) –8% yr ⁻¹	(Singh et al., 2018)
	2009-2019	Japan (Fukue Island) –5.8 ± 1.5% yr ⁻¹	(Kanaya et al., 2020)
	2009-2015	India (Darjeeling mountain site) –5% yr ⁻¹	(Sarkar et al., 2019)
OA	2001-2015	USA (IMPROVE sites east of 100°W) –2% yr ⁻¹	(Malm et al., 2017)
Total Carbon (EC+OC)	1990-2010	USA (IMPROVE sites) Western USA: –4 to –5% yr ⁻¹ Eastern USA: –1 to –2% yr ⁻¹	(Hand et al., 2013)
	2002-2010	Spain (Montseny rural site) –5% yr ⁻¹	(Querol et al., 2013)

1
2 **[END TABLE 6.6 HERE]**
3
4

5 Knowledge of carbonaceous aerosol atmospheric abundance continues to rely on global models due to a lack
6 of global scale observations. For BC, models agree within a factor of two with measured surface mass
7 concentrations in Europe and North America, but underestimate concentrations at the Arctic surface by one
8 to two orders of magnitude, especially in winter and spring (Lee et al., 2013a; Lund et al., 2018a). For OA,
9 AeroCom models underestimate surface mass concentrations by a factor 2 over urban areas, as their low
10 horizontal resolution prevents them from resolving local pollution peaks (Tsigaridis et al., 2014; Lund et al.,
11 2018a). Models agree within a factor of two with OA surface concentrations measured at remote sites, where
12 surface concentrations are more spatially uniform (Tsigaridis et al., 2014).
13

14 Lifetimes in models are estimated to 5.5 days ±35% for BC and 6.0 days ±29% for OA (median ± 1 standard
15 deviation) according to an ensemble of 14 models (Gliß et al., 2021). Disagreement in simulated lifetime
16 leads to horizontal and vertical variations in predicted carbonaceous aerosol concentrations, with
17 implications for radiative forcing (Lund et al., 2018b) (Samset et al., 2013). Airborne campaigns have
18 provided valuable vertical profile measurements of carbonaceous aerosol concentrations (Schwarz et al.,
19 2013; Freney et al., 2018; Hodgson et al., 2018; Morgan et al., 2019; Schulz et al., 2019; Zhao et al., 2019a).
20 Compared to those measurements, models tend to transport BC too high in the atmosphere, suggesting that
21 lifetimes are not larger than 5.5 days (Samset et al., 2013; Lund et al., 2018b). Newly developed size-
22 dependent wet scavenging parameterisation for BC (Taylor et al., 2014; Schroder et al., 2015; Ohata et al.,
23 2016; Zhang et al., 2017a; Ding et al., 2019; Moteki et al., 2019; Motos et al., 2019) may lead to decreased
24 BC lifetimes and improve agreement with observed vertical profiles.
25

26 Simulated BC burdens show a large spread among models (Gliß et al., 2021), despite using harmonised
27 primary emissions, because of differences in BC removal efficiency linked to different treatment of ageing
28 and mixing, particularly in strong source regions. The multi-model median BC burden for the year 2010
29 from Gliß et al. (2021), based on 14 AeroCom models, is 0.131 ± 0.047 Tg (median ± standard deviation).
30 That range encompasses values reported by independent single model estimates (Huang et al., 2013; Lee et
31 al., 2013b; Sharma et al., 2013b; Wang et al., 2014a; Tilmes et al., 2019).
32

33 Simulated OA burdens also show a large spread among global models, with Gliß et al. (2021) reporting a
34 multi-model median of 1.91 ± 0.65 Tg for the year 2010. This large spread reflects the wide range in the
35 complexity of the OA parameterizations, particularly for SOA formation, as well as in the primary OA
36 emissions (Tsigaridis et al., 2014; Gliß et al., 2021). The uncertainties are particularly large in model
37 estimates of SOA production rates, which vary between 10 and 143 Tg yr⁻¹ (Tsigaridis et al., 2014; Hodzic et

1 al., 2016; Tilmes et al., 2019). While the level of complexity in the representation of OA in global models
2 has increased since AR5 (Shrivastava et al., 2017; Hodzic et al., 2020) limitations in process level
3 understanding of the formation, aging and removal of organic compounds lead to uncertainties in the global
4 model predictions of global OA burden and distribution as well as the relative contribution of POA and SOA
5 to OA. Jo et al. (2016) estimated that BrC contributes about 20% of total OA burden. That would give BrC a
6 burden similar to that of BC (*low confidence*), enhancing the overall forcing exerted by carbonaceous aerosol
7 absorption (Zhang et al., 2020).

8
9 In summary, the lack of global scale observations of carbonaceous aerosol, its complex atmospheric
10 chemistry, and the large spread in its simulated global budget and burdens means that there is only *low*
11 *confidence* in the quantification of the present-day atmospheric distribution of individual components of
12 carbonaceous aerosols. Global trends in carbonaceous aerosols cannot be characterised due to limited
13 observations, but sites representative of background conditions have reported multi-year declines in BC over
14 several regions of the Northern Hemisphere.

15 16 17 **6.3.6 Implications of SLCF abundances for Atmospheric Oxidizing Capacity**

18
19 The atmospheric oxidising capacity is determined primarily by tropospheric hydroxyl (OH) radical and to a
20 smaller extent by NO₃ radical, ozone, hydrogen peroxide (H₂O₂) and halogen radicals. OH is the main sink
21 for many SLCFs, including methane, halogenated compounds (HCFCs and HFCs), CO and NMVOCs,
22 controlling their lifetimes and consequently their abundance and climate influence. OH initiated oxidation of
23 methane, CO and NMVOCs in the presence of NO_x leads to the production of tropospheric ozone. OH also
24 contributes to the formation of aerosols from oxidation of SO₂ to sulphate and VOCs to secondary organic
25 aerosols. The evolution of atmospheric oxidising capacity of the Earth driven by human activities and natural
26 processes is, therefore, of significance for climate and air quality concerns.

27
28 The main source of tropospheric OH is the photoexcitation of tropospheric ozone that creates an
29 electronically excited oxygen atom which reacts with water vapour producing OH. A secondary source of
30 importance for global OH is the recycling of peroxy radicals formed by the reaction of OH with reduced and
31 partly oxidized species, including methane, CO and NMVOCs. In polluted air, NO_x emissions control the
32 secondary OH production, while in pristine air it occurs via other mechanisms involving, in particular,
33 isoprene (Lelieveld et al., 2016; Wennberg et al., 2018). Knowledge of the effect of isoprene oxidation on
34 OH recycling has evolved tremendously over the past decade facilitating mechanistic explanation of elevated
35 OH concentrations observed in locations characterised by low NO_x levels (Hofzumahaus et al., 2009; Paulot
36 et al., 2009; Peeters et al., 2009, 2014; Fuchs et al., 2013). Since AR5, inclusion of improved chemical
37 mechanisms in some CTMs suggest advances in understanding of the global OH budget, however these
38 improvements have yet to be incorporated in CMIP6-generation ESMS.

39
40 As a result of the complex photochemistry, tropospheric OH abundance is sensitive to changes in SLCF
41 emissions as well as climate. Increases in methane, CO, NMVOCs reduce OH while increases in water
42 vapour and temperature, incoming solar radiation, NO_x and tropospheric ozone enhance OH. The OH level
43 thus responds to climate change and climate variability via its sensitivity to temperature and water vapour as
44 well as the influence of climate on natural emissions (e.g., wetland methane emissions, lightning NO_x,
45 BVOCs, fire emissions) with consequent feedbacks on climate (Section 6.4.5). Climate modes of variability,
46 like ENSO, also contribute to OH variability via changes in lightning NO_x emissions and deep convection
47 (Turner et al., 2018), and fire emissions (Rowlinson et al., 2019).

48
49 Global scale OH observations are non-existent because of its extremely short lifetime (~1 s) and therefore
50 global OH abundance and its time variations are either inferred from atmospheric measurements of methyl
51 chloroform (MCF) (Prinn et al., 2018 and references therein) or derived from global atmospheric chemistry
52 models (Lelieveld et al., 2016). AR5 reported small interannual OH variations in the 2000s based on
53 atmospheric inversions of MCF observations (within ±5%) and global CCMs and CTMs (within ±3%) (Ciais
54 et al., 2013).

1 Since AR5, there is much closer agreement in the estimates of interannual variations in global mean OH
2 derived from atmospheric inversions, empirical reconstruction, and global CCMs and ESMs with an estimate
3 of 2 to 3% over the 1980 to 2015 period (Table 6.7). While the different methodologies agree on the
4 occurrence of small inter-annual variations, there is much debate over longer term global OH trend. Two
5 studies using multi-box model inversions of MCF and methane observations suggest large positive and
6 negative trends since 1990s in global mean OH (Rigby et al., 2017; Turner et al., 2017), however, both find
7 that observational constraints are weak such that a wide range of multi-annual OH variations are possible.
8 Indeed, (Naus et al., 2019) find an overall positive global OH trend over the past two decades (Table 6.7)
9 after accounting for uncertainties and biases in atmospheric MCF and CH₄ inversions confirming the
10 weakness in observational constraints for deriving OH trends. Global ESMs, CCMs and CTMs exhibit
11 stabilized or increasing global OH after 1980 contrary to the lack of trend derived from some atmospheric
12 inversions and empirical reconstructions (Table 6.7). In particular, a three member ensemble of ESMs
13 participating in the AerChemMIP/CMIP6 agrees that global OH has increased since 1980 by around 9%
14 (Figure 6.9) with an associated reduction in methane lifetime (Stevenson et al., 2020). This positive OH
15 trend is in agreement with the ~7% OH increase derived by assimilating global-scale satellite observations of
16 CO over the 2002-2013 period (with CO declining trends) into a CCM (Gaubert et al., 2017; see section
17 6.3.4). Multi-model sensitivity analysis suggests that increasing OH since 1980 is predominantly driven by
18 changes in anthropogenic SLCF emissions with complementary influence of increasing NO_x and decreasing
19 CO emissions (Stevenson et al., 2020).

20
21 Over paleo time scales, proxy-based observational constraints from CH₄ and formaldehyde suggest
22 tropospheric OH to be a factor of 2 to 4 lower in the last glacial maximum (LGM) relative to preindustrial
23 levels, though these estimates are highly uncertain (Alexander and Mickley, 2015). Global models, in
24 contrast, exhibit no change in tropospheric OH (and consequently in methane lifetime) at the LGM relative
25 to the preindustrial (Murray et al., 2014; Quiquet et al., 2015), however the sign and magnitude of OH
26 changes are sensitive to model predictions of changes in natural emissions, including lightning NO_x and
27 BVOCs, and model representation of isoprene oxidation chemistry (Achakulwisut et al., 2015; Hopcroft et
28 al., 2017).

29
30 Regarding change since preindustrial era, at the time of the AR5, ensemble mean of 17 global models
31 participating in ACCMIP indicated little change in tropospheric OH from 1850 to 2000. This was due to the
32 competing and finally offsetting changes in factors enhancing or reducing OH with a consequent small
33 decline in methane lifetime (Naik et al., 2013; Voulgarakis et al., 2013). However, there was large diversity
34 in both the sign and magnitude of past OH changes across the individual models attributed to disparate
35 implementation of chemical and physical processes (Nicely et al., 2017; Wild et al., 2020). Analysis of
36 historical simulations from three CMIP6 ESMs indicates little change in global mean OH from 1850 to about
37 1980 (Stevenson et al., 2020). However, there is no observational evidence of changes in global OH since
38 1850 up to early 1980s to evaluate the ESMs.

39
40 In summary, global mean tropospheric OH does not show a significant trend from 1850 up to around 1980
41 (*low confidence*). There is conflicting information from global models constrained by emissions versus
42 observationally constrained inversion methods over the 1980-2014 period. A positive trend since 1980
43 (about 9 % increase over 1980-2014) is a robust feature among ESMs and CCMs and there is *medium*
44 *confidence* that this trend is mainly driven by increases in global anthropogenic NO_x emissions and decreases
45 in CO emissions. There is *limited evidence* and *medium agreement* for positive trends or absence of trends
46 inferred from observation-constrained methods. Overall, there is *medium confidence* that global mean OH
47 has remained stable or exhibited a positive trend since the 1980s.

1 [START TABLE 6.7 HERE]

2
3 **Table 6.7:** Summary of global OH trends and interannual variability from studies post 2010.

4

Reference	Time period	OH trends and IAV	Approach
Inversion and empirical methods based on observations			
(Montzka et al., 2011)	1998-2007	2.3±1.3% (IAV)	3D inversion
(Ciais et al., 2013)	2000s	<±5% (IAV)	AR5 based on inversions
(McNorton et al., 2016)	1993-2011	±2.3% (IAV)	Box model inversion
(Rigby et al., 2017)	1980-2014	10% increase from the late 1990s to 2004; 10% decrease from 2004 to 2014	Box model inversion
(Turner et al., 2017)	1983-2015	~7% increase in 1991-2001; 7% decrease in 2003-2016	Box model inversion
(Nicely et al., 2018)	1980-2015	1.6% (IAV)	Empirical reconstruction
(McNorton et al., 2018)	2003-2015	1.8±0.4% decrease	3D inversion
(Naus et al., 2019)	1994-2015	3.8±3.2% increase	Box model inversion
(Patra et al., 2021)	1996-2015	2-3% IAV, no trend	3D inversion
Global CTMs, CCMs and ESMs			
(John et al., 2012)	1860-2005 1980-2000	6% decrease ~3% increase	CCM
(Holmes et al., 2013)	1997-2009	0.7–1.1% (IAV)	Multi-model CTMs
(Ciais et al., 2013)	2000s	<±3 % (IAV)	AR5 based on CCMs
(Murray et al., 2013)	1998–2006	increasing trend	3D CTM
(Naik et al., 2013)	1980-2000 1850-2000	3.5±2.2 % increase −0.6 ± 8.8%	Multi-model CCMs/CTMs
(Murray et al., 2014)	1770s-1990s	5.3% increase	CCM
(Dalsøren et al., 2016)	1970-2012	8% increase	3D CTM
(Gaubert et al., 2017)	2002-2013	7% increase	CCM with assimilated satellite CO observations
(Zhao et al., 2019b)	1960-2010 1980-2000	1.9 ± 1.2 % (IAV) 4.6 ± 2.4 % increase	Multimodel CCMs/CTMs
(Stevenson et al., 2020)	1980-2014 1850-1980	9% increase no trend	Multimodel ESMs

5
6 [END TABLE 6.7 HERE]

7
8
9 [START FIGURE 6.9 HERE]

10
11 **Figure 6.9:** Time evolution of global annual mean tropospheric hydroxyl (OH) over the historical period, expressed as a percentage anomaly relative to the mean over 1998-2007. a) Results from three CMIP6 models, including UKESM1-0LL (green), GFDL-ESM4 (blue), and CESM2-WACCM (red), are shown; the shaded light green and light red bands show mean over multiple ensemble members for UKESM1-0LL (3) and CESM2-WACCM (3) models, respectively with the multi-model mean anomalies shown in thick black line. b) multimodel mean OH anomalies for 1980-2015 period compared with those derived

1 from observational-based inversions from (Montzka et al., 2011; Rigby et al., 2017; Turner et al., 2017;
2 Nicely et al., 2018; Naus et al., 2019; Patra et al., 2021) in the zoomed box. Further details on data
3 sources and processing are available in the chapter data table (Table 6.SM.1).
4

5 **[END FIGURE 6.9 HERE]**
6

8 **6.4 SLCF radiative forcing and climate effects**

9

10 The radiative forcing on the climate system introduced by SLCFs is distinguished from that of LLGHGs by
11 the diversity of forcing mechanisms for SLCFs, and the challenges of constraining these mechanisms via
12 observations and of inferring their global forcings from available data. Chapter 7 assesses the global
13 estimates of ERFs due to SLCF abundance changes. This section assesses the characteristics (e.g., spatial
14 patterns, temporal evolution) of forcings, climate response and feedbacks due to SLCFs relying primarily on
15 results from CMIP6 models. Additionally, the ERFs for several aerosol-based forms of solar radiation
16 modification (SRM) are discussed in Section 6.4.6.
17

18 Forcing and climate response due to changes in SLCFs are typically estimated from global models that vary
19 in their representation of the various chemical, physical, and radiative processes (see Box 6.1) affecting the
20 causal chain from SLCF emissions to climate response (Figure 6.2). The AR5 noted that the representation
21 of aerosol processes varied greatly in CMIP5 models and that it remained unclear what level of
22 sophistication is required to properly quantify aerosol effects on climate (Boucher et al., 2013). Since the
23 AR5, (Ekman, 2014) found that the CMIP5 models with the most complex representations of aerosol
24 impacts on cloud microphysics had the largest reduction in biases in surface temperature trends. CMIP6-
25 generation CCMs that simulate aerosol and cloud size distributions better represent the effect of a volcanic
26 eruption on lower atmosphere clouds than a model with aerosol-mass only (Malavelle et al., 2017). This
27 highlights the need for skillful simulation of conditions underlying aerosol-cloud interactions, such as the
28 distribution, transport and properties of aerosol species, in addition to the interactions themselves (see
29 Chapter 7). In advance of CMIP6, representations of aerosol processes and aerosol-cloud interactions in
30 ESMs have generally become more comprehensive (Gliß et al., 2021; Meehl et al., 2020; Thornhill et al.,
31 2021b; see also Section 1.5), with enhanced links to aerosol emissions and gas-phase chemistry. Many
32 CMIP6 models (see Annex II Table AII.5) now simulate aerosol number size distribution, in addition to
33 mass distribution, which is a prerequisite for accurately simulating number concentrations of cloud
34 condensation nuclei (CCN) (Bellouin et al., 2013) while some CMIP6 models use prescribed aerosol optical
35 properties to constrain aerosol forcing (e.g., Stevens et al., 2017). Hence, the range of complexity in aerosol
36 modeling noted in CMIP5 is still present in the CMIP6 ensemble. Although simulated CCN have been
37 compared to surface (Fanourgakis et al., 2019) and aircraft (Reddington et al., 2017) measurements, with
38 mixed results, the lack of global coverage limits confidence in the evaluations. Evaluations of aerosol optical
39 depths have been more wide ranging (section 6.3.5; Gliß et al., 2021) but are less relevant to aerosol-cloud
40 interactions as they do not allow to evaluate vertical profiles, aerosol-cloud overlap regions, aerosol type or
41 number. Nevertheless, biases in simulated patterns and trends in aerosol optical depths, alongside biases in
42 cloud fractions (Vignesh et al., 2020), likely affect quantifications of the aerosol-cloud interactions.
43

44 In summary, CMIP6 models generally represent more processes that drive aerosol-cloud interactions than the
45 previous generation of climate models, but there is only *medium confidence* that those enhancements
46 improve their fitness for the purpose of simulating radiative forcing of aerosol-cloud interactions because
47 only a few studies have identified the level of sophistication required to do so. In addition, the challenge of
48 representing the small-scale processes involved in aerosol-cloud interactions, and a lack of relevant model-
49 data comparisons, does not allow to quantitatively assess the progress of the models from CMIP5 to CMIP6
50 in simulating the underlying conditions relevant for aerosol-cloud interactions at this time.
51

53 **6.4.1 Historical Estimates of Regional Short-lived Climate Forcing**

54

55 The highly heterogeneous distribution of SLCF abundances (section 6.3) translates to strong heterogeneity in

1 the spatial pattern and temporal evolution of forcing and climate responses due to SLCFs. This section
2 assesses the spatial patterns of the current forcing due to aerosols and their historical evolution by region.

3
4 In AR5, the confidence in the spatial patterns of aerosol and ozone forcing was lower than that for the global
5 mean because of the large spread in the regional distribution simulated by global models, and was assessed
6 as ‘medium’. The AR5 assessment was based on aerosol and ozone RFs and aerosol ERFs (with fixed SSTs)
7 from ACCMIP and a small sample of CMIP5 experiments (Myhre et al., 2013b; Shindell et al., 2013). For
8 this assessment, the spatial distribution of aerosol ERF (ERF_{AER}) due to human-induced changes in aerosol
9 concentrations over 1850-2014 is quantified based on results from a 7 member ensemble of CMIP6 ESMS
10 including interactive gas and aerosol chemistry analysed in AerChemMIP. There is insufficient information
11 to estimate the spatial patterns of ozone ERF from CMIP6, however, the spatial patterns in SLCF ERF are
12 dominated by that from aerosol ERF over most regions (e.g., Shindell et al., 2015). The ERF_{AER} includes
13 contributions from both direct aerosol-radiation (ERF_{dir}) and indirect aerosol-cloud interactions (ERF_{aci}) (see
14 Section 7.3.3) and is computed as the difference between radiative fluxes from simulations with
15 time-evolving aerosol and their precursor emissions and identical simulations but with these emissions held at
16 their 1850 levels (Collins et al., 2017). Both the simulations are driven by time-evolving SSTs and sea ice
17 from the respective coupled model historical simulation and therefore differ from ERFs computed using
18 fixed pre-industrial SST and sea-ice fields (Section 7.3.1), but the effect of this difference is generally small
19 (Forster et al., 2016). A correction for land surface temperature change (Section 7.3.1) is not available from
20 these data to explicitly quantify the contribution from adjustments. The ESMS included here used the CMIP6
21 anthropogenic and biomass burning emissions for ozone and aerosol precursors but varied in their
22 representation of natural emissions, chemistry and climate characteristics contributing to spread in the
23 simulated concentrations (also see section 6.3) and resulting forcings, partly reflecting uncertainties in the
24 successive processes (Thornhill et al., 2021b).

25
26 The geographical distribution of the ensemble mean ERF_{AER} over the 1850-2014 period is highly
27 heterogeneous (Figure 6.10a) in agreement with AR5. Negative ERF_{AER} is greatest over and downwind of
28 most industrialized regions in the Northern Hemisphere and to some extent over tropical biomass burning
29 regions, with robust signals. Largest negative forcing occurs over East Asia and South Asia, followed by
30 Europe and North America, reflecting the changes in anthropogenic aerosol emissions in the recent decades
31 (Section 6.2). Positive ERF_{AER} over high albedo areas, including cryosphere, deserts and clouds, also found
32 in AR5 and attributed to absorbing aerosols, are not robust across the small CMIP6 ensemble applied here.
33 Regionally aggregated shortwave (SW) and longwave (LW) components of the ERF_{AER} exhibit similar large
34 variability across regions (Figure 6.10b). The SW flux changes come from aerosol-radiation and aerosol-
35 cloud interactions while the small positive LW flux changes come from aerosol-cloud interactions (related to
36 liquid water path changes (Section 7.3.2.2)). These spatial patterns in ERF_{AER} are similar to the patterns
37 reported in AR5.

38
39 Time evolution of 20 year means of regional net ERF_{AER} shows that the regions are divided into two groups
40 depending on whether the mean ERF_{AER} attains its negative peak value in the 1970s to 1980s (e.g., Europe,
41 North America) or in the late 1990s to 2000s (e.g., Asia, South America) (Figure 6.11). Qualitatively, this
42 shift in the distribution of ERF_{AER} trends is consistent with the regional long-term trends in aerosol precursor
43 emissions (Section 6.21; Figures 6.18 and 6.19) and their abundances (Section 6.3). However, at finer
44 regional scales, there are regions where sulphate aerosols are still following an upward trend (e.g., South
45 Asia; see Section 6.3.5) implying that the trends in ERF_{AER} may not have shifted for these regions. The
46 continental scale ERF_{AER} trends are also in line with the satellite-observed AOD trends assessed in Section
47 2.2.6. Global mean ERF_{AER} reaches maximum negative values in the mid-1970s and its magnitude gradually
48 decreases thereafter. This weakening of the negative forcing since 1990 agrees with findings by who
49 attribute this to a reduction in global mean SO_2 emissions combined with an increase in global BC (Myhre et
50 al., 2017). Uncertainties in model simulated aerosol ERF distribution and trends can result from intermodel
51 variations in the representation of aerosol-cloud interactions and aerosol microphysical processes as also
52 demonstrated by Bauer et al., (2020).

53
54 In summary, the spatial and temporal distribution of the net aerosol ERF from 1850 to 2014 is highly
55 heterogeneous (*high confidence*). Globally, there has been a shift from increase to decrease of the negative

1 net aerosol ERF driven by trends in aerosol and their precursor emissions (*high confidence*). However, the
2 timing of this shift varies by continental-scale region and has not occurred for some finer regional scales.
3
4

5 **[START FIGURE 6.10 HERE].**

6
7 **Figure 6.10: Multi-model mean Effective radiative forcings (ERFs) due to aerosol changes between 1850 and**
8 **recent-past (1995-2014).** Panel (a) shows the spatial distribution of the net ERF with area-weighted
9 global mean ERF shown at the lower right corner. Uncertainty is represented using the advanced
10 approach: No overlay indicates regions with robust signal, where $\geq 66\%$ of models show change greater
11 than variability threshold and $\geq 80\%$ of all models agree on sign of change; diagonal lines indicate regions
12 with no change or no robust signal, where $< 66\%$ of models show a change greater than the variability
13 threshold; crossed lines indicate regions with conflicting signal, where $\geq 66\%$ of models show change
14 greater than variability threshold and $< 80\%$ of all models agree on sign of change. For more information
15 on the advanced approach, please refer to the Cross-Chapter Box Atlas.1. Panel (b) shows the mean
16 shortwave and longwave ERF for each of the 14 regions defined in the Atlas. Violins in panel (b) show
17 the distribution of values over regions where ERFs are significant. ERFs are derived from the difference
18 between top of the atmosphere (TOA) radiative fluxes for Aerosol Chemistry Model Intercomparison
19 Project (AerChemMIP) experiments histSST and histSST-piAer (Collins et al., 2017) averaged over
20 1995-2014 (Box 1.4, Chapter 1). The results come from 7 Earth System Models: MIROC6, MPI-I-ESM-
21 1-2-HAM, MRI-ESM2-0, GFDL-ESM4, GISS-E2-1-G, NorESM2-LM, and UKESM-0-LL. These data
22 can be seen in the Interactive Atlas. Further details on data sources and processing are available in the
23 chapter data table (Table 6.SM.1).
24

25 **[END FIGURE 6.10 HERE]**

26
27
28 **[START FIGURE 6.11 HERE]**

29
30 **Figure 6.11: Time evolution of 20-year multi-model mean averages of the annual area-weighted mean regional**
31 **net Effective Radiative Forcings (ERFs) due to aerosols.** Each of the 14 major regions in the Atlas are
32 shown, as well as the global mean, using the models and model experiments as in Figure 6.10. Further
33 details on data sources and processing are available in the chapter data table (Table 6.SM.1).
34

35 **[END FIGURE 6.11 HERE]**

36 37 38 **6.4.2 Emission-based Radiative Forcing and effect on GSAT**

39
40 The ERFs attributable to emissions versus concentrations for several SLCFs including ozone and methane
41 are different. A concentration change, used to assess the abundance-based ERF, results from the changes in
42 emissions of multiple species and subsequent chemical reactions. The corollary is that the perturbation of a
43 single emitted compound can induce subsequent chemical reactions and affect the concentrations of several
44 climate forcers (chemical adjustments), this is what is accounted for in emission-based ERF. Due to non-
45 linear chemistry (cf. Section 6.3) and non-linear aerosol-cloud interactions (section 7.3.3.2), the ERF
46 attributed to the individual species cannot be precisely defined, and can only be estimated through model
47 simulations. For example, the ERF attributed to methane emissions, which includes indirect effects through
48 ozone formation and oxidation capacity with feedbacks on the methane lifetime, depend non-linearly on the
49 concentrations of NO_x , CO and VOCs. This means that the results from the model simulations depend to
50 some extent on the chosen methodology. In AR5 (based on Shindell et al., 2009; Stevenson et al., 2013) the
51 attribution was done by removing the anthropogenic emissions of individual species one-by-one from a
52 control simulation for present day conditions. Further, only the radiative forcings, and not the ERF (mainly
53 including the effect of aerosol-cloud interactions) were attributed to the emitted compounds.
54

55 Since AR5, the emission estimates have been revised and extended for CMIP6 (Hoesly et al., 2018), the
56 models have been further developed, the period has been extended (1750-2019, vs 1750-2011 in AR5) and
57 the experimental setup for the model simulations has changed (Collins et al., 2017) making a direct

1 comparison of results difficult. Figure 6.13 shows the global and annual mean ERF attributed to emitted
2 compounds over the period 1750-2019 based on AerChemMIP simulations (Thornhill et al., 2021b) where
3 anthropogenic emissions or concentrations of individual species were perturbed from 1850 levels to 2014
4 levels (methodology described in Supplementary Material 6.SM.1).

5
6
7 **[START FIGURE 6.12 HERE]**

8
9 **Figure 6.12:** Contribution to effective radiative forcing (ERF) (left) and global mean surface air temperature (GSAT)
10 change (right) from component emissions between 1750 to 2019 based on CMIP6 models (Thornhill et
11 al., 2021b). ERFs for the direct effect of well-mixed greenhouse gases (WMGHGs) are from the
12 analytical formulae in section 7.3.2, H₂O (strat) is from Table 7.8. ERFs for other components are multi-
13 model means from Thornhill et al. (2021b) and are based on ESM simulations in which emissions of one
14 species at a time are increased from 1850 to 2014 levels. The derived emission-based ERFs are rescaled
15 to match the concentration-based ERFs in Figure 7.6. Error bars are 5-95% and for the ERF account for
16 uncertainty in radiative efficiencies and multi-model error in the means. ERF due to aerosol-radiation
17 (ERF_{ari}) and cloud effects are calculated from separate radiation calls for clear-sky and aerosol free
18 conditions (Ghan, 2013; Thornhill et al., 2021b). “Cloud” includes cloud adjustments (semi-direct effect)
19 and indirect aerosol-cloud interactions (ERF_{aci}). The aerosol components (SO₂, organic carbon, black
20 carbon) are scaled to sum to -0.22 W m⁻² for ERF_{ari} and -0.84 W m⁻² for “cloud” (section 7.3.3). For
21 GSAT estimates, time series (1750-2019) for the ERFs have been estimated by scaling with
22 concentrations for WMGHGs and with historical emissions for SLCFs. The time variation of ERF_{aci} for
23 aerosols is from Chapter 7. The global mean temperature response is calculated from the ERF time series
24 using an impulse response function (see Cross-Chapter Box 7.1) with a climate feedback parameter of -
25 1.31 W m⁻² °C⁻¹. Further details on data sources and processing are available in the chapter data table
26 (Table 6.SM.1).

27
28 **[END FIGURE 6.12 HERE]**

29
30
31 The ERF based on primary CO₂ emissions is slightly lower than the abundance-based estimate (section
32 7.3.2.1) because the abundance-based ERF combines the effect of primary CO₂ emissions and a small
33 additional secondary contribution from atmospheric oxidation of CH₄, CO, and VOCs (4%) of fossil origin,
34 consistent with AR5 findings.

35
36 Ozone depleting substances, such as N₂O and halocarbons, cause a reduction in stratospheric ozone, which
37 affects ozone and OH production in the troposphere through radiation UV changes (and thus affect methane).
38 They also have indirect effects on aerosols and clouds (Karset et al., 2018) since changes in oxidants induce
39 changes in the oxidation of aerosol precursors.

40
41 The net ERF from N₂O emissions is estimated to be 0.24 (0.13 to 0.34) W m⁻², which is very close to the
42 abundance-based estimate of 0.21 W m⁻² (Section 7.3.2.3). The indirect contributions from N₂O are
43 relatively minor with negative (methane lifetime) and positive (ozone and clouds) effects nearly
44 compensating each other. Emissions of halogenated compounds, including CFCs and HCFCs, were assessed
45 as very likely causing a net positive ERF in the AR5. However recent studies (Morgenstern et al., 2020;
46 O’Connor et al., 2021; Thornhill et al., 2021b) find strong adjustments in Southern Hemisphere aerosols and
47 clouds such that the *very likely* range in the emission based ERF for CFC+HCFCs+HFCs now also include
48 negative values.

49
50 For methane emissions, in addition to their direct effect, there are indirect positive ERFs from methane
51 enhancing its own lifetime, causing ozone production, enhancing stratospheric water vapor, and influencing
52 aerosols and the lifetimes of HCFCs and HFCs (Myhre et al., 2013b; O’Connor et al., 2021). The ERF from
53 methane emissions is considerably higher than the ERF estimate resulting from its abundance change. The
54 central estimate with the *very likely* range is 1.21 (0.90 to 1.51) W m⁻² for emission-based estimate versus
55 0.54 W m⁻² for abundance-based estimate (cf. section 7.3.5). The abundance-based ERF estimate for CH₄
56 results from contributions of its own emissions and the effects of several other compounds, some decreasing
57 CH₄ lifetime, notably NO_x, which importantly reduce the CH₄ abundance-based ERF. Emissions of CO and

1 NMVOCs both indirectly contribute to a positive ERF through enhancing ozone production in the
2 troposphere and increasing the methane lifetime. For CO and NMVOCs of fossil origin there is also a 0.07
3 W m^{-2} contribution to CO_2 from their oxidation. The *very likely* total ERF of CO and NMVOCs emissions is
4 estimated to 0.45 (0.22 to 0.67) W m^{-2} .

5
6 NO_x causes a positive ERF through enhanced tropospheric ozone production and a negative ERF through
7 enhanced OH concentrations that reduce the methane lifetime. There is also a small negative ERF
8 contribution through formation of nitrate aerosols, although only 3 of the AerChemMIP models include
9 nitrate aerosols. The best estimate of the net ERF from changes in anthropogenic NO_x emissions is -0.29 (-
10 0.57 to 0.0) W m^{-2} . The magnitude is somewhat greater than the AR5 estimate (-0.15 (-0.34 to + 0.02) W m^{-2})
11 but with similar level of uncertainty. The difference between AR6 and AR5 estimates is possibly due to
12 the different modeling protocols (see Supplementary Material 6.SM.1).

13
14 Anthropogenic emissions of SO_2 lead to formation of sulphate aerosols and a negative ERF through aerosol-
15 radiation and aerosol-cloud interactions. The emission-based ERF_{aci} , which was not previously considered in
16 AR5, is now included. The estimated ERF is thus considerably more negative than the AR5 estimate with a
17 radiative forcing of -0.4 W m^{-2} , despite the decline of ERF due to aerosols since 2011 (Figure 6.12, Section
18 7.3.3.1.3). SO_2 emissions are estimated to contribute to a negative ERF of -0.90 (-0.24 to -1.56) W m^{-2} , with
19 -0.22 W m^{-2} from aerosol-radiation interactions and -0.68 W m^{-2} from aerosol-cloud interactions. Emissions
20 of NH_3 lead to formation of ammonium-nitrate aerosols with an estimated ERF of -0.03 W m^{-2} .

21
22 The best estimate for the ERF due to emissions of BC is reduced from the AR5, and is now estimated to be
23 0.063 (-0.28 to 0.42) W m^{-2} with an uncertainty also including negative values. As discussed in Section
24 7.3.3.1.2, a significant portion of the positive BC forcing from aerosol-radiation interactions is offset by
25 negative atmospheric adjustments due to cloud changes as well as lapse rate and atmospheric water vapor
26 changes, resulting in a smaller positive net ERF for BC compared with AR5. The large range in the forcing
27 estimate stems from variation in the magnitude and sign of atmospheric adjustments across models and is
28 related to the differences in the model treatment of different processes affecting BC (e.g., ageing, mixing)
29 and its interactions with clouds and cryosphere (Thornhill et al., 2021b and Section 7.3.3.). The emission-
30 based ERF for organic carbon aerosols is -0.20 (-0.03 to -0.41) W m^{-2} , a weaker estimate compared with
31 AR5 attributed to stronger absorption by OC (section 7.3.3.1.2).

32
33 The emission-based contributions to GSAT change (Figure 6.12) were not assessed in AR5, but with the
34 ERF from aerosol-cloud interactions attributed to the emitted compounds there is now a better foundation for
35 this assessment. The contribution to emissions-based ERF at 2019 (left panel in figure 6.12) is scaled by the
36 historical emissions (over the period 1750 to 2019) of each compound to reconstruct the historical time series
37 of ERF. An impulse response function (see Cross-Chapter Box 7.1 and Supplementary Material 7.SM5.2) is
38 then applied to obtain the contribution of SLCF emissions to the GSAT response. Due to the non-linear
39 chemical and physical processes described above relating emissions to ERF, and the additional non-linear
40 relations between ERF and GSAT, these emission-based estimates of GSAT responses strongly depend on
41 the methodology applied to estimate ERF and GSAT (see Supplementary Material 6.SM.2). Therefore, the
42 relative contribution of each compound through its primary emissions versus secondary formation or
43 destruction (e.g. for methane emissions its ozone versus methane contributions), by construction (omitting
44 the non-linear processes), will be equal for ERF and GSAT. Uncertainties in the GSAT response are
45 estimated using the assessed range of the ECS from Chapter 7 of this report. For most of the SLCFs the
46 uncertainty in the GSAT response is dominated by the uncertainty in the relationship between emissions and
47 the ERF.

48
49 The contributions from the emitted compounds to GSAT broadly follow their contributions to the ERF,
50 mainly because their evolution over the past decades have been relatively similar and slow enough compared
51 to their lifetimes to be reflected similarly in their ERF and GSAT despite the delay of the GSAT response to
52 ERF changes (see Section 6.6.1). However, for some SLCFs, e.g. SO_2 , that have been reduced globally, their
53 contribution to GSAT change is slightly higher compared with that of CO_2 than their relative contribution to
54 ERF because the peak in their ERF change has already occurred (see Section 6.4.1) whereas the peak of their
55 GSAT effect started to decline recently (see also Figure 7.9). This is due to the inertia of the climate system

1 delaying the full response of GSAT to a change in forcing (Figure 6.15).

2
3 In summary, emissions of SLCFs, especially CH₄, NO_x and SO₂, have substantial effects on effective
4 radiative forcing (ERF) (*high confidence*). The net global emissions-based ERF of NO_x is negative and that
5 of NMVOCs is positive, in agreement with the AR5 assessment (*high confidence*). For methane, the
6 emission-based ERF is twice as high as the abundance-based ERF (*high confidence*). SO₂ emissions make
7 the dominant contribution to the ERF associated with the aerosol–cloud interaction (*high confidence*).
8 The contributions from the emitted compounds to GSAT broadly follow their contributions to the ERF (*high*
9 *confidence*). However, due to the inertia of the climate system delaying the full GSAT response to a change
10 in forcing, the contribution to GSAT change due to SO₂ emission is slightly higher compared with that due to
11 CO₂ emissions (than their relative contributions to ERF) because the peak in emission-induced SO₂ ERF has
12 already occurred.

13 14 15 **6.4.3 Climate responses to SLCFs**

16
17 This section briefly discusses the climate response to SLCFs, in particular to changes in aerosols, and gathers
18 complementary information and assessments from Chapters 3, 7, 8, and 10.

19
20 Warming or cooling atmospheric aerosols, such as BC and sulphate, can affect temperature and precipitation
21 in distinct ways by modifying the shortwave and longwave radiation, the lapse rate of the troposphere, and
22 influencing cloud microphysical properties (see Box 8.1 and Section 10.1.4.1.4). An important distinction
23 between scattering and absorbing aerosols is the opposing nature of their influences on circulation, clouds,
24 and precipitation, besides surface temperature as evident from the contrasting regional climate responses to
25 regional aerosol emissions (e.g., Lewinschal et al., 2019; Sand et al., 2020; also see Chapters 8 and 10).

26
27 On the global scale, as assessed in Chapter 3, anthropogenic aerosols have *likely* cooled GSAT since 1850–
28 1900 driven by the negative aerosol forcing, while it is *extremely likely* that human-induced stratospheric
29 ozone depletion has primarily driven stratospheric between 1979 and the mid-1990s. Multiple modelling
30 studies support the understanding that present-day emissions of SO₂, precursor for sulphate aerosols, are the
31 dominant driver of near-surface air temperature in comparison to BC or OC even though, for some regions,
32 BC forcing plays a key role (Baker et al., 2015; Samset et al., 2016; Stjern et al., 2017; Zanis et al., 2020).
33 Further, there is *high confidence* that the aerosol-driven cooling has led to detectable large-scale water cycle
34 changes since at least the mid-20th century as assessed in Chapter 8. The overall effect of surface cooling
35 from anthropogenic aerosols is to reduce global precipitation and alter large-scale atmospheric circulation
36 patterns (*high confidence*), primarily driven by the cooling effects of sulphate aerosols (Section 8.2.1). In
37 addition, there is *high confidence* that darkening of snow through the deposition of black carbon and other
38 light absorbing particles enhances snow melt (SROCC Chapter 3, see also Section 7.3.4.3).

39
40 In AR5, there was low confidence in the overall understanding of climate response to spatially varying
41 patterns of forcing, though there was medium to high confidence in some regional climate responses, such as
42 the damped warming of the NH and shifting of the ITCZ from aerosols, and positive feedbacks enhancing
43 the local response from high-latitude snow and ice albedo changes. Since AR5, the relationship between
44 inhomogeneous forcing and climate response is better understood providing further evidence of the climate
45 influence of SLCFs (aerosols and ozone in particular) on global to regional scales (Collins et al., 2013;
46 Shindell et al., 2015; Aamaas et al., 2017; Kasoar et al., 2018; Persad and Caldeira, 2018; Wilcox et al.,
47 2019) which differ from the relatively homogeneous spatial influence from LLGHGs.

48
49 Large geographical variations in aerosol ERFs (section 6.4.1) affect global and regional temperature
50 responses (Shindell et al., 2015, Myhre et al., 2013a). A multimodel CMIP6 ensemble mean results (Figure
51 6.13) show cooling over almost all areas of the globe in response to increases of aerosol and their precursor
52 emissions from 1850 to recent past (1995–2014). While the ERF has hotspots, the temperature response is
53 more evenly distributed in line with the results of CMIP5 models including the temperature response to
54 ozone changes (Shindell et al 2015). The ensemble mean global mean surface temperature decreases by
55 0.66 ± 0.51 °C while decreasing by 0.97 ± 0.54 °C for Northern Hemisphere and 0.34 ± 0.2 °C for Southern

1 Hemisphere. The zonal mean temperature response is negative at all latitudes (high confidence) and becomes
2 more negative with increasing latitude, with a maximum ensemble mean decrease of around 2.7 °C at
3 Northern polar latitudes. The zonal mean response is not directly proportional to the zonal mean forcing,
4 especially in the Arctic where the temperature response is cooling while the local ERF is positive (6.10).
5 This is consistent with prior studies showing that the Arctic, in particular, is highly sensitive to forcing at NH
6 midlatitudes (e.g., Sand et al., 2013b; Shindell and Faluvegi, 2009) and with results from CMIP5 models
7 (Shindell et al., 2015) (more on Arctic below). Thus, there is *high confidence* that the temperature response
8 to aerosols is more asymmetric than the response to WMGHGs and negative at all latitudes.

9
10 The asymmetric aerosol and greenhouse gas forcing on regional scale climate responses have also been
11 assessed to lead to contrasting effects on precipitation in Chapter 8. The asymmetric historical radiative
12 forcing due to aerosols led to a southward shift in the tropical rain belt (*high confidence*) and contributed to
13 the Sahel drought from the 1970s to the 1980s (*high confidence*). Furthermore, the asymmetry of the forcing
14 led to contrasting effects in monsoon precipitation changes over West Africa, South Asia and East Asia over
15 much of the mid-20th century due to GHG-induced precipitation increases counteracted by anthropogenic
16 aerosol-induced decreases (*high confidence*) (see Section 8.3, Box 8.1).

17
18 The Arctic region is warming considerably faster than the rest of the globe (Atlas 5.9.2.2) and, generally,
19 studies indicate that this amplification of the temperature response toward the Arctic has important
20 contribution from local and remote aerosol forcing (Stjern et al., 2017; Westervelt et al., 2018). Several
21 studies indicate that changes in long-range transport of sulphate and BC from northern midlatitudes can
22 potentially explain a significant fraction of Arctic warming since 1980s (e.g., Breider et al., 2017; Navarro et
23 al., 2016; Ren et al., 2020). Modeling studies show that changes in midlatitude aerosols have influenced
24 Arctic climate by changing the radiative balance through aerosol-radiation and aerosol-cloud interactions and
25 enhancing poleward heat transport (Navarro et al., 2016; Ren et al., 2020b). Idealized aerosol perturbation
26 studies have shed further light on the sensitivity of Arctic temperature response to individual aerosol species.
27 Studies show relatively large responses in the Arctic to BC perturbations and reveal the importance of
28 remote BC forcing by rapid adjustments (Sand et al., 2013a; Stjern et al., 2017; Liu et al., 2018; Yang et al.,
29 2019a). Perturbations in SO₂ emissions over major emitting regions in the Northern Hemisphere have been
30 shown to produce largest Arctic temperature responses (Kasoar et al., 2018; Lewinschal et al., 2019).

31
32 The effects of changes in aerosols on local and remote changes in temperature, circulation and precipitation
33 are sensitive to a number of model uncertainties affecting aerosol sources, transformation, and resulting
34 climate effects. Therefore, regional climate effects in global model studies must be interpreted with caution.
35 When investigating the climate response to regional aerosol emissions, such uncertainties are likely to be
36 confounded even further by the variability between models in regional climate and circulation patterns,
37 leading to greater intermodel spread at regional scales than at a global scale (Baker et al., 2015; Kasoar et al.,
38 2016).

39
40 In summary, over the historical period, changes in aerosols and their ERF have primarily contributed to cool
41 the surface temperature partly masking the human-induced warming (*high confidence*). Radiative forcings
42 induced by aerosol changes lead to both local and remote changes in temperature (*high confidence*). The
43 temperature response preserves hemispheric asymmetry of the ERF but is more latitudinally uniform with
44 strong amplification of the temperature response towards the Arctic (*medium confidence*).

45
46
47 **[START FIGURE 6.13 HERE]**

48
49 **Figure 6.13: Multi-model mean surface air temperature response due to aerosol changes between 1850 and**
50 **recent-past (1995-2014).** Calculation is based on the difference between CMIP6 ‘historical’ and
51 AerChemMIP ‘hist-piAer’ experiments, where a) is the spatial pattern of the annual mean surface air
52 temperature response, and b) is the mean zonally averaged response. Model means are derived from years
53 1995-2014. Uncertainty is represented using the advanced approach: No overlay indicates regions with
54 robust signal, where ≥66% of models show change greater than variability threshold and ≥80% of all
55 models agree on sign of change; diagonal lines indicate regions with no change or no robust signal, where
56 <66% of models show a change greater than the variability threshold; crossed lines indicate regions with

1 conflicting signal, where $\geq 66\%$ of models show change greater than variability threshold and $< 80\%$ of all
2 models agree on sign of change. For more information on the advanced approach, please refer to the
3 Cross-Chapter Box Atlas.1. AerChemMIP models MIROC6, MRI-ESM2-0, NorESM2-LM, GFDL-
4 ESM4, GISS-E2-1-G, UKESM1-0-LL are used in the analysis. Further details on data sources and
5 processing are available in the chapter data table (Table 6.SM.1).
6

7 **[END FIGURE 6.13 HERE]**
8
9

10 **6.4.4 Indirect radiative forcing through effects of SLCFs on the carbon cycle**

11
12 Deposition of reactive nitrogen (Nr) (i.e. NH_3 and NO_x) increases the plant productivity and carbon
13 sequestration in N-limited forests and grasslands, and also in open and coastal waters and open ocean. Such
14 inadvertent fertilization of the biosphere can lead to eutrophication and reduction in biodiversity in terrestrial
15 and aquatic ecosystems. AR5 assessed that it is *likely* that Nr deposition over land currently increases natural
16 CO_2 sinks, in particular in forests, but the magnitude of this effect varies between regions (Ciais et al., 2013).
17 Increasing Nr deposition or the synergy between increasing Nr deposition and atmospheric CO_2
18 concentration could have contributed to the increasing global net land CO_2 sink [Section 5.2.1.4.1].
19

20 Ozone uptake itself damages photosynthesis and reduces plant growth with consequences for the carbon and
21 water cycles (Ainsworth et al., 2012; Emberson et al., 2018). AR5 concluded there was robust evidence of
22 the effect of ozone on plant physiology and subsequent alteration of the carbon storage but considered
23 insufficient quantification of and a lack of systematic incorporation of the ozone effect in carbon-cycle
24 models as a limitation to assess the terrestrial carbon balance (Ciais et al., 2013). Since AR5 several more
25 ESMs have incorporated interactive ozone-vegetation damage resulting in an increase in evidence to support
26 the influence of ozone on the land carbon cycle. The new modelling studies tend to focus on ozone effects on
27 plant productivity rather than the land carbon storage and agree that ozone-induced gross-primary
28 productivity (GPP) losses are largest today in eastern USA, Europe and eastern China ranging from 5-20%
29 on the regional scale (Yue and Unger, 2014; Lombardozzi et al., 2015; Yue et al., 2017; Oliver et al., 2018)
30 (*low confidence*). There is *medium evidence* and *high agreement* based on observational studies and models
31 that ozone-vegetation interactions further influence the climate system, including water and carbon cycles by
32 affecting stomatal control over plant transpiration of water vapour between the leaf surface and atmosphere
33 (Arnold et al., 2018; Hoshika et al., 2015; Lombardozzi et al., 2013; Sun et al., 2012; VanLoocke et al.,
34 2012; Wittig et al., 2007). While some modelling studies suggest that the unintended Nr deposition
35 fertilization effect in forests may potentially offset the ozone-induced carbon losses (Felzer et al., 2007; de
36 Vries et al., 2017), complex interactions have been observed between ozone and Nr deposition to ecosystems
37 that have not yet been included in ESMs. For some plants, the effects of increasing ozone on root biomass
38 become more pronounced as Nr deposition increased, and the beneficial effects of Nr on root development
39 were lost at higher ozone treatments (Mills et al., 2016). Reducing uncertainties in ozone vegetation damage
40 effects on the carbon cycle requires improved information on the sensitivity of different plant species to
41 ozone, and measurements of ozone dose-response relationships for tropical plants, which are currently
42 lacking. Surface ozone effect on the land carbon sink and indirect CO_2 forcing, therefore, remains uncertain.
43 Collins et al. (2010) showed that adding in the effects of surface ozone on vegetation damage and reduced
44 uptake of CO_2 added about 10 % to the methane emission metrics and could change the sign of the NO_x
45 metrics. However, this estimate has to be considered as an upper limit due to limitation of the
46 parameterization used by Sitch et al. (2007) considering more recent knowledge and is thus not included in
47 the current metrics (Section 7.6.1.3).
48

49 Tropospheric aerosols influence the land and ocean ecosystem productivity and the carbon cycle through
50 changing physical climate and meteorology (Jones, 2003; Cox et al., 2008; Mahowald, 2011; Unger et al.,
51 2017) and through changing deposition of nutrients including nitrogen, sulphur, iron and phosphorous
52 (Mahowald et al., 2017; Kanakidou et al., 2018). There is *high evidence* and *high agreement* from field
53 (Oliveira et al., 2007; Cirino et al., 2014; Rap et al., 2015; Wang et al., 2018c) and modelling (Mercado et
54 al., 2009; Strada and Unger, 2016; Lu et al., 2017; Yue et al., 2017) studies that aerosols affect plant
55 productivity through increasing the diffuse fraction of downward shortwave radiation although the

1 magnitude and importance to the global land carbon sink is controversial. At large-scales the dominant effect
2 of aerosols on the carbon cycle is *likely* a global cooling effect of the climate (Jones, 2003; Mahowald, 2011;
3 Unger et al., 2017) (*medium confidence*). We assess that these interactions between aerosols and the carbon
4 cycle are currently too uncertain to constrain quantitatively the indirect CO₂ forcing.
5

6 In summary, reactive nitrogen, ozone and aerosols affect terrestrial vegetation and carbon cycle through
7 deposition and effects on large scale radiation (*high confidence*) but the magnitude of these effects on the
8 land carbon sink, ecosystem productivity and indirect CO₂ forcing remain uncertain due to the difficulty in
9 disentangling the complex interactions between the effects. As such we assess it to be of second order in
10 comparison to the direct CO₂ forcing (*high confidence*) but, at least for ozone, it could add a substantial
11 (positive) forcing compared with its direct forcing (*low confidence*).
12
13

14 6.4.5 Non-CO₂ biogeochemical feedbacks

15
16 Climate change-induced changes in atmospheric composition and forcing due to perturbations in natural
17 processes constitute an Earth system feedback amplifying (positive feedback) or diminishing (negative
18 feedback) the initial climate perturbation (Ciais et al., 2013; Heinze et al., 2019). Quantification of these
19 biogeochemical feedbacks is important to allow for better estimate of the expected effects of emission
20 reduction policies for mitigating climate change and the effect on the allowable global carbon budget (Lowe
21 and Bernie, 2018). Biogeochemical feedbacks due to changes in carbon-cycle are assessed in Section 5.4.5,
22 while physical and biophysical climate feedbacks are assessed in Section 7.4.2. Additionally, non-CO₂
23 biogeochemical feedbacks due to climate driven changes in methane sources and N₂O sources and sinks are
24 assessed in Section 5.4.7. The goal of this section is to estimate the feedback parameter (α as defined in
25 section 7.4.1.1) from climate-induced changes in atmospheric abundances or lifetimes of SLCFs mediated by
26 natural processes or atmospheric chemistry. These non-CO₂ biogeochemical feedbacks act on time scales of
27 years to decades and have important implications for climate sensitivity and emission abatement policies.
28 The feedback parameter is quantified entirely from ESMs that expand the complexity of CCMs by coupling
29 the physical climate and atmospheric chemistry to land and ocean biogeochemistry. In AR5, α for non-CO₂
30 biogeochemical feedbacks was estimated from an extremely limited set of modeling studies with much less
31 confidence associated with the estimate. Since AR5, ESMs have advanced to include more feedback
32 processes facilitating a relatively more robust assessment of α . CMIP6 ESMs participating in AerChemMIP
33 performed coordinated sets of experiments (Collins et al., 2017) facilitating the consistent estimation of α
34 (Thornhill et al., 2021a) and we rely on this multi-model analysis for the best estimates (Table 6.9).
35 Considering the consistent methodology, the assessed central values and 5-95% ranges for α are based on the
36 AerChemMIP estimates. The full range of model uncertainty is not captured in AerChemMIP because of the
37 relatively small ensemble size therefore estimates from studies using other models or with different protocols
38 are discussed to reinforce or critique these values.
39

40 **Climate-Sea-spray feedback:** Sea-spray emissions from ocean surfaces influence climate directly or
41 indirectly through the formation of CCN as discussed in Section 6.2.1.2. They are sensitive to sea-surface
42 temperature, sea ice extent, as well as wind speed and are therefore expected to feedback on climate
43 (Struthers et al., 2013). However, there are large uncertainties in the strength of climate feedback from sea-
44 spray aerosols because of the diversity in the model representation of emissions (many represent sea-salt
45 emissions only) and their functional dependence on environmental factors noted above, in situ atmospheric
46 chemical and physical processes affecting the sea-spray lifetime, and aerosol-cloud interactions (Struthers et
47 al., 2013; Soares et al., 2016; Nazarenko et al., 2017). Additional work is needed to identify how sea-spray
48 and POA emissions respond to shifts in ocean biology and chemistry in response to warming, ocean
49 acidification, and changes in circulation patterns (Burrows et al., 2018) and affect CCN and INP formation
50 (DeMott et al., 2016). AerChemMIP models, representing only the sea-salt emissions, agree that the
51 sea-salt-climate feedback is negative, however there is a large range in the feedback parameter indicating
52 large uncertainties (Table 6.9).
53

54 **Climate-DMS feedback:** Dimethyl sulphide (DMS) is produced by marine phytoplankton and is emitted to
55 the atmosphere where it can lead to the subsequent formation of sulphate aerosol and CCN (see Section

1 6.2.1.2). Changes in DMS emissions from ocean could feedback on climate through their response to
2 changes in temperature, solar radiation, ocean mixed-layer depth, sea-ice extent, wind-speed, nutrient
3 recycling or shifts in marine ecosystems due to ocean acidification and climate change, or atmospheric
4 processing of DMS into CCN (Heinze et al., 2019). Models with varying degrees of representation of the
5 relevant biogeochemical processes and effects on DMS fluxes produce diverging estimates of changes in
6 DMS emission strength under climate change resulting in large uncertainties in the DMS–sulphate–cloud
7 albedo feedback (Bopp et al., 2004; Kloster et al., 2007; Gabric et al., 2013). In AR5, the climate–DMS
8 feedback parameter was estimated to be $-0.02 \text{ W m}^{-2} \text{ }^{\circ}\text{C}^{-1}$ based on a single model. Since AR5, new
9 modeling studies using empirical relationships between pH and total DMS production find that global DMS
10 emissions decrease due to combined ocean acidification and climate change leading to a strong positive
11 climate feedback (Six et al., 2013; Schwinger et al., 2017). However, another study argues for a much
12 weaker positive feedback globally due to complex and compensating regional changes in marine ecosystems
13 (Wang et al., 2018b). The AerChemMIP multimodel analysis suggests small positive feedback (Table 6.9),
14 consistent with these recent studies, but with large uncertainties in the magnitude of α .

15
16 **Climate-dust feedback:** Mineral dust is the most abundant aerosol type in the atmosphere, when considering
17 aerosol mass, and affects the climate system by interacting with both longwave and shortwave radiation as
18 well as contributing to the formation of CCN and INP. Because dust emissions are sensitive to climate
19 variability (e.g., through changes in the extent of arid land) (Section 6.2.1.2), it has been hypothesized that
20 the climate-dust feedback could be an important feedback loop in the climate system. Since AR5, an
21 improved understanding of the shortwave absorption properties of dust as well as a consensus that dust
22 particles are larger than previously thought has led to a revised understanding that the magnitude of radiative
23 forcing due to mineral dust is small (Kok et al., 2017; Ryder et al., 2018). A recent study notes that global
24 models underestimate the amount of coarse dust in the atmosphere and accounting for this limitation raises
25 the possibility that dust emissions warm the climate system (Adebiyi and Kok, 2020). Model predictions of
26 dust emissions in response to future climate change range from an increase (Woodward et al., 2005) to a
27 decrease (Mahowald and Luo, 2003), thus leading to high uncertainties on the sign of the climate-dust
28 feedback. Since the AR5, Kok et al. (2018) estimated the direct dust-climate feedback parameter, from
29 changes in the dust direct radiative effect only, to be in the range -0.04 to $+0.02 \text{ W m}^{-2} \text{ }^{\circ}\text{C}^{-1}$. The assessed
30 central value and the 5-95% range of climate-dust feedback parameter based on AerChemMIP ensemble
31 (Table 6.9) is within the range of the published estimate, however both the magnitude and sign of α are
32 model-dependent.

33
34 **Climate-ozone feedback:** Changes in ozone concentrations in response to projected climate change have
35 been shown to lead to a potential climate-atmospheric chemistry feedback. Chemistry-climate models
36 consistently project a decrease in lower tropical stratospheric ozone levels due to enhanced upwelling of
37 ozone poor tropospheric air associated with surface warming driven strengthening of the Brewer-Dobson
38 circulation (Bunzel and Schmidt, 2013). Further, models project an increase in middle and extratropical
39 stratospheric ozone due to increased downwelling through the strengthened Brewer-Dobson circulation
40 (Bekki et al., 2013; Dietmüller et al., 2014). These stratospheric ozone changes induce a net negative global
41 mean ozone radiative feedback (Dietmüller et al., 2014). Tropospheric ozone shows a range of responses to
42 climate with models generally agreeing that warmer climate will lead to decreases in the tropical lower
43 troposphere owing to increased water vapour and increases in the sub-tropical to mid-latitude upper
44 troposphere due to increases in lightning and stratosphere-to-troposphere transport (Stevenson et al., 2013).
45 A small positive feedback is estimated from climate-induced changes in global mean tropospheric ozone
46 (Dietmüller et al., 2014) while a small negative feedback is estimated by (Heinze et al., 2019) based on the
47 model results of (Stevenson et al., 2013). Additionally, these ozone feedbacks induce a change in
48 stratospheric water vapor amplifying the feedback due to stratospheric ozone (Stuber et al., 2001). Since
49 AR5, several modeling studies have estimated the intensity of meteorology driven ozone feedbacks on
50 climate from either combined tropospheric and stratospheric ozone changes or separately with contrasting
51 results. One study suggests no change (Marsh et al., 2016), while other studies report reductions of
52 equilibrium climate sensitivity ranging from 7-8% (Dietmüller et al., 2014; Muthers et al., 2014) to 20%
53 (Nowack et al., 2015). The estimate of this climate-ozone feedback parameter is very strongly model
54 dependent with values ranging from -0.13 to $-0.01 \text{ W m}^{-2} \text{ }^{\circ}\text{C}^{-1}$ though there is agreement that it is negative.
55 The assessed central value and the 5-95% range of climate-ozone feedback parameter based on

1 AerChemMIP ensemble is within the range of these published estimates but closer to the lower bound. This
2 climate-ozone feedback factor does not include the feedback on ozone from lightning changes which is
3 discussed separately below.

4
5 **Climate-BVOC feedback:** BVOCs, such as isoprene and terpenes, are produced by land vegetation and
6 marine plankton (Section 6.2.1.2). Once in the atmosphere, BVOCs and their oxidation products lead to the
7 formation of secondary organic aerosols (SOA) exerting a negative forcing, and increased ozone
8 concentrations and methane lifetime exerting a positive forcing. BVOC emissions are suggested to lead to a
9 climate feedback in part because of their strong temperature dependence observed under present-day
10 conditions (Kulmala et al., 2004; Arneth et al., 2010a). Their response to future changes in climate and CO₂
11 levels remains uncertain (see Section 6.2.2.3). Estimates of the climate-BVOC feedback parameter are
12 typically based on global models which vary in their level of complexity of emissions parameterization,
13 BVOC speciation, the mechanism of SOA formation and the interaction with ozone chemistry (Thornhill et
14 al., 2021a). Since AR5, observational studies (Paasonen et al. 2013) and models (Scott et al. 2018) estimate
15 the feedback due to biogenic SOA (via changes in BVOC emissions) to be in the range of about -0.06
16 to $-0.01 \text{ W m}^{-2} \text{ }^{\circ}\text{C}^{-1}$. The assessed central estimate of the climate-BVOC feedback parameter based on the
17 AerChemMIP ensemble suggests that climate-induced increases in SOA from BVOCs will lead to a strong
18 cooling effect that will outweigh the warming from increased ozone and methane lifetime, however the
19 uncertainty is large (Thornhill et al., 2021a).

20
21 **Climate-Lightning NO_x feedback:** As discussed in Section 6.2.1.2, climate change influences lightning NO_x
22 emissions. Increases in lightning NO_x emissions will not only increase tropospheric ozone and decrease
23 methane lifetime but also increase the formation of sulphate and nitrate aerosols, via oxidant changes,
24 offsetting the positive forcing from ozone. The response of lightning NO_x to climate change remains
25 uncertain and is highly dependent on the parameterization of lightning in ESMs (Finney et al., 2016b; Clark
26 et al., 2017) (also see section 6.2.1.2). AerChemMIP multi-model ensemble mean estimate a net negative
27 climate feedback from increases in lightning NO_x in a warming world (Thornhill et al., 2021a). All
28 AerChemMIP models use a cloud-top height lightning parameterization that predicts increases in lightning
29 with warming. However, a positive climate-lightning NO_x feedback cannot be ruled out because of the
30 dependence of the response to lightning parameterizations as discussed in section 6.2.2.1.

31
32 **Climate-CH₄ Lifetime feedback:** Warmer and wetter climate will lead to increases in OH and oxidation rates
33 leading to reduced atmospheric methane lifetime – a negative feedback (Naik et al., 2013; Voulgarakis et al.,
34 2013). Furthermore, since OH is in turn removed by methane, the climate-methane lifetime feedback will be
35 amplified (Prather, 1996) (also see section 6.3.1). Based on the multimodel results of Voulgarakis et al.
36 (2013), α for climate-methane lifetime is estimated to be $-0.030 \pm 0.01 \text{ W m}^{-2} \text{ }^{\circ}\text{C}^{-1}$ by Heinze et al., (2019).
37 The assessed central value of α based on the AerChemMIP ensemble is within the range of this estimate but
38 with greater uncertainty (Thornhill et al., 2021a).

39
40 **Climate-Fire feedback:** Wildfires are a major source of SLCF emissions (section 6.2.2.6). Climate change
41 has the potential to enhance fire activity (see Sections 5.4.3.2, Chapter 12.4) thereby enhancing SLCF
42 emissions leading to feedbacks. Climate driven increases in fire could potentially lead to offsetting feedback
43 from increased ozone and decreased methane lifetime (due to increases in OH) leaving the feedback from
44 aerosols to dominate with an uncertain net effect (e.g., Landry et al., 2015). AR5 assessment of climate-fire
45 feedbacks included a value of α due to fire aerosols to be in the range of -0.03 to $+0.06 \text{ W m}^{-2} \text{ }^{\circ}\text{C}^{-1}$ based on
46 Arneth et al. (2010). A recent study estimates climate feedback due to fire aerosols to be greater than that
47 due to BVOCs, with a value of α equal to -0.15 (-0.24 to -0.05) $\text{W m}^{-2} \text{ }^{\circ}\text{C}^{-1}$ (Scott et al., 2018). Clearly, the
48 assessment of fire related non-CO₂ biogeochemical feedbacks is very uncertain because of limitations in the
49 process understanding of the interactions between climate, vegetation and fire dynamics, and atmospheric
50 chemistry and their representation in the current generation ESMs. Some AerChemMIP ESMs include the
51 representation of fire dynamics but do not activate their interaction with atmospheric chemistry. Given the
52 large uncertainty and lack of information from AerChemMIP ESMs, we do not include a quantitative
53 assessment of climate-fire feedback for AR6.

54
55 In summary, climate-driven changes in emissions, atmospheric abundances or lifetimes of SLCFs are

1 assessed to have an overall cooling effect, i.e., a negative feedback parameter of $-0.20 \text{ W m}^{-2} \text{ }^{\circ}\text{C}^{-1}$ with a 5-
 2 95% range of -0.41 to $+0.01 \text{ W m}^{-2} \text{ }^{\circ}\text{C}^{-1}$, thereby reducing climate sensitivity (see Section 7.4.2.5.1). This
 3 net feedback parameter is obtained by summing the assessed estimates for the individual feedback given in
 4 Table 6.8. *Confidence* in the magnitude and the sign of most of the individual as well as the total non-CO₂
 5 biogeochemical feedbacks remains *low* as evident from the large range in the value of α . This large
 6 uncertainty is attributed to the diversity in model representation of the relevant chemical and biogeochemical
 7 processes based on limited process-level understanding.

8
 9
 10 [START TABLE 6.8 HERE]

11
 12 **Table 6.8:** Assessed central estimates and the *very likely* ranges (5-95%) of non-CO₂ biogeochemical feedback
 13 parameter (α_x) based on the AerChemMIP ensemble estimates (Thornhill et al., 2021a). As in Section
 14 7.4.1.1, α_x ($\text{W m}^{-2} \text{ }^{\circ}\text{C}^{-1}$) 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
 15 energy balance in response to a change in x induced by a change in surface temperature (T). The 5-95%
 16 range is calculated as mean \pm standard deviation * 1.645 for each feedback. The level of confidence in
 17 these estimates is *low* owing to the large model spread. Published estimates of α are also shown for
 18 comparison.
 19

Non-CO ₂ Biogeochemical Climate Feedback (x)	# of AerChemMIP Models	Assessed Central Estimate and Very Likely Range of Feedback Parameter (α_x) $\text{W m}^{-2} \text{ }^{\circ}\text{C}^{-1}$	Published estimates of α_x $\text{W m}^{-2} \text{ }^{\circ}\text{C}^{-1}$
Sea-salt	6	-0.049 (-0.13 to +0.03)	-0.08 (Paulot et al., 2020)
DMS	3	0.005 (0.0 to 0.01)	-0.02 (Ciais et al., 2013)
Dust	6	-0.004 (-0.02 to +0.01)	-0.04 to +0.02 (Kok et al., 2018)
Ozone	4	-0.064 (-0.08 to -0.04)	-0.015 (Dietmüller et al., 2014), -0.06 (Muthers et al., 2014, stratospheric ozone changes only), -0.01 (Marsh et al., 2016, stratospheric ozone changes only), -0.13 (Nowack et al., 2015, stratospheric ozone and water vapor changes), -0.007 ± 0.009 (Heinze et al., 2019, tropospheric ozone changes only)
BVOC	4	-0.05 (-0.22 to +0.12)	-0.06 (Scott et al., 2017, aerosol effects only), -0.01 (Paasonen et al., 2013; indirect aerosol effects only), 0-0.06 (Ciais et al., 2013)
Lightning	4	-0.010 (-0.04 to +0.02)	
Methane lifetime	4	-0.030 (-0.12 to +0.06)	-0.30 ± 0.01 (Heinze et al., 2019)
Total non-CO ₂ Biogeochemical Feedbacks assessed in this chapter		-0.200 (-0.41 to +0.01)	0.0 ± 0.15 (Sherwood et al., 2020)

20
 21 [END TABLE 6.8 HERE]
 22
 23

24 6.4.6 ERF by aerosols in proposed Solar Radiation Modification

25
 26 Solar radiation modification (SRM, see also Sections 4.6.3.3, 8.6.3) have the potential to exert a significant
 27 ERF on the climate, mainly by affecting the SW component of the radiation budget (e.g. Caldeira et al.,
 28 2013; Lawrence et al., 2018; National Academy of Sciences, 2015). The possible ways and the extent to

1 which the most commonly discussed options may affect radiative forcing is addressed in this section. Side-
2 effects of SRM on stratospheric ozone and changes in atmospheric transport due to radiative heating of the
3 lower stratosphere are discussed in Section 4.6.3.3.

4
5 Stratospheric aerosol injections have the potential to achieve a high negative global ERF, with maximum
6 ERFs ranging from -2 to -5 W m⁻² (Niemeier and Timmreck, 2015; Weisenstein et al., 2015; Niemeier and
7 Schmidt, 2017; Kleinschmitt et al., 2018). The magnitude of the maximum achievable ERF depends on the
8 chosen aerosol type and mixture, internal structure and size, or precursor gas (e.g. SO₂), as well as the
9 injection strategy; latitude, altitude, magnitude, and season of injections, plume dispersal, model
10 representation of aerosol microphysics, and ambient aerosol concentrations (Rasch et al., 2008; Robock et
11 al., 2008; Pierce et al., 2010; Weisenstein et al., 2015; Laakso et al., 2017; Macmartin et al., 2017; Dai et al.,
12 2018; Kleinschmitt et al., 2018; Vattioni et al., 2019; Visoni et al., 2019). For sulphur, the radiative forcing
13 efficiency is of around -0.1 to -0.4 W m⁻² / (TgS yr⁻¹) (Niemeier and Timmreck, 2015; Weisenstein et al.,
14 2015; Niemeier and Schmidt, 2017). Different manufactured aerosols, such as ZrO₂, TiO₂, and Al₂O₃, have
15 different ERF efficiencies compared to sulphate (Ferraro et al., 2011; Weisenstein et al., 2015; Dykema et
16 al., 2016; Jones et al., 2016). The aerosol size distribution influences the optical properties of an aerosol
17 layer, and hence the ERF efficiency, which also depends on the dispersion, transport, and residence time of
18 the aerosols.

19
20 For marine cloud brightening (MCB), seeded aerosols may affect both cloud microphysical and
21 macrophysical properties (see also Section 7.3.3.2). By principle, MCB relies on ERF_{aci} through the so-called
22 Twomey effect (Twomey, 1977), but ERF_{ari} may be of equal magnitude as shown in studies that consider
23 spraying of sea salt outside tropical marine cloud areas (Jones and Haywood, 2012; Partanen et al., 2012;
24 Alterskjær and Kristjánsson, 2013; Ahlm et al., 2017). The maximum negative ERF estimated from
25 modelling is within the range of -0.8 to -5.4 W m⁻² (Latham et al., 2008; Rasch et al., 2009; Jones et al.,
26 2011; Partanen et al., 2012; Alterskjær and Kristjánsson, 2013). For dry sea salt, the ERF efficiency is
27 estimated to be within the range of -3 to -10 W m⁻² / (Pg yr⁻¹), when emitted over tropical oceans in ESMs in
28 the Geoengineering Intercomparison Project (GeoMIP) (Ahlm et al., 2017). Cloud resolving models reveal
29 complex behaviour and response of stratocumulus clouds to seeding, in that the ERF efficiency depends on
30 meteorological conditions, and the ambient aerosol composition, where lower background particle
31 concentrations may increase the ERF_{aci} efficiency (Wang et al., 2011). Seeding could suppress precipitation
32 formation and drizzle, and hence increase the lifetime of clouds, preserving their cooling effect (Ferek et al.,
33 2000). In contrast, cloud lifetime could be decreased by making the smaller droplets more susceptible to
34 evaporation. Modelling studies have shown that a positive ERF_{aci} (warming) could also result from seeding
35 clouds with too large aerosols (Pringle et al., 2012; Alterskjær and Kristjánsson, 2013). These processes, and
36 the combination of these, are not well understood, and may have a limited representation in models, or
37 counteracting errors (Mülmenstädt and Feingold, 2018), lending *low to medium confidence* to the ERF
38 estimates.

39
40 Modelled ERF_{aci} associated with cirrus cloud thinning (CCT) cover a wide range in the literature, and the
41 maximum are of the order of -0.8 to -3.5 W m⁻², though they are of *low confidence*, with some studies using
42 more simplified representations (Mitchell and Finnegan, 2009; Storelvmo et al., 2013; Kristjánsson et al.,
43 2015; Jackson et al., 2016; Muri et al., 2018; Gasparini et al., 2020). ERF_{aci} for CCT is mainly affected by
44 particle seeding concentrations, with an optimum around 20 L⁻¹, according to limited evidence from models
45 (Storelvmo et al., 2013). Seeding leading to higher particle concentrations could lead to a warming
46 (Storelvmo et al., 2013; Penner et al., 2015; Gasparini and Lohmann, 2016). The lack of representation of
47 processes related to, for example, heterogeneous and homogeneous freezing and their prevalence, is a
48 dominant source of uncertainty in ERF estimates, in addition to less research activity.

49
50 In summary, the aerosol and cloud microphysics involved with SRM are not well understood, with a varying
51 degree of lack of representation of relevant processes in models. ERF of up to several W m⁻² is reported in
52 the literature, with SAI at the higher end and CCT with lower potentials, though it remains a challenge to
53 establish ERF potentials and efficacies with confidence. Modelling studies have been published with more
54 sophisticated treatment of SRM since AR5, but the uncertainties, such as cloud-aerosol radiation
55 interactions, remain large (*high confidence*).

6.5 Implications of changing climate on AQ

Air pollutants can be impacted by climate change through physical changes affecting meteorological conditions, chemical changes affecting their lifetimes, and biological changes affecting their natural emissions (Kirtman et al., 2013). Changes in meteorology affect air quality directly through modifications of atmospheric transport patterns (e.g., occurrence and length of atmospheric blocking episodes, ventilation of the polluted boundary layer), extent of mixing layer and stratosphere-troposphere exchange (STE) for surface ozone (Von Schneidemesser et al., 2015), and through modifications of the rate of reactions that generate secondary species in the atmosphere. Changing precipitation patterns in a future climate also influence the wet removal efficiency, in particular for atmospheric aerosols (Hou et al., 2018). Processes at play in non-CO₂ biogeochemical feedbacks (Section 6.4.5) are also involved in the perturbation of atmospheric pollutants (see Section 6.2.2).

This section relies on observational studies performed by analysing correlation between specific meteorological conditions projected to occur more frequently in the future and surface pollutants, and global and regional scale modelling studies considering solely climate change in the future. We also assess the surface ozone and PM_{2.5} changes based on CMIP6 models analysed in AerChemMIP, considering climate change in isolation with emissions in 2050 from SSP3-7.0 scenario (see Section 6.7.1). Air quality being highly variable in space and time, the use of regional atmospheric chemistry models is necessary to characterise the effect of future climate on air quality properly. However, difficulties for such assessment arise from the need for long simulations that include complex chemistry-natural system interactions with high computational cost, in addition to the difficulty related to the regionalisation of climate change (see Section 10.3.1.2). Changes in the occurrence of weather patterns influencing air pollution (e.g., anticyclonic stagnation conditions, transport pathways from pollution sources, convection) due to climate change are assessed in Chapters 4 and 11.

6.5.1 Effect of climate change on surface O₃

AR5 assessed with high confidence that (Kirtman et al., 2013), in unpolluted regions, higher water vapour abundances and temperatures in a warmer climate would enhance ozone chemical destruction, leading to lower baseline⁵ surface ozone levels. In polluted regions, AR5 assessed with medium confidence that higher surface temperatures will trigger regional feedbacks in chemistry and local emissions that will increase surface ozone and intensity of surface O₃ peaks.

The response of surface ozone to climate induced Earth system changes is complex due to counteracting effects. Studies considering the individual effects of climate driven changes in specific precursor emissions or processes show increases in surface ozone under warmer atmosphere for some processes. This is indeed the case for enhanced STE and stratospheric ozone recovery (Sekiya and Sudo, 2014; Banerjee et al., 2015; Hess et al., 2015; Meul et al., 2018; Morgenstern et al., 2018; Akritidis et al., 2019) or the increase of soil NO_x emissions (Wu et al., 2008; Romer et al., 2018), which can each lead to 1 to 2 ppb increase in surface ozone. Other processes, in particular deposition or those related to emissions from natural systems (see section 6.2.2) are expected to play a key role in future surface ozone and even occurrence of pollution events (e.g. in the case of wildfires) but their effects are difficult to quantify in isolation.

Since the AR5, several studies have investigated the net effect of climate change on surface ozone, based on either global or regional model projections. A systematic and quantitative comparison of the ozone change, however, is difficult due to the variety of models with different complexities in the representation of natural emissions, chemical mechanisms, and physical processes as well as the surface O₃ metrics applied for analysis. Processes like temperature-, CO₂-sensitive BVOC emissions, deposition, and branching ratio in isoprene nitrates chemistry have been shown to be particularly sensitive (Squire et al., 2015; Val Martin et al., 2015; Schnell et al., 2016; Pommier et al., 2018). More robust protocol are now used more commonly

⁵ Baseline ozone is defined as the observed ozone at a site when it is not influenced by recent, locally emitted or anthropogenically produced pollution (Jaffe et al., 2018).

1 comprising, notably, longer simulations necessary to separate change from interannual variability
2 (Lacressonnière et al., 2016; Garcia-Menendez et al., 2017) (Barnes et al., 2016). However, the amplitude of
3 climate change penalty on ozone over polluted regions may be different in high-resolution (regional and
4 urban-scale) models in comparison to coarse resolution global models as a number of controlling processes
5 are resolution-dependent including e.g local emissions, sensitivity to the chemical regime (VOC limited
6 versus NO_x limited) (Lauwaet et al., 2014; Markakis et al., 2014, 2016).

7
8 Consistent with AR5 findings, global surface ozone concentration decreases by up to -1.2 to 2.3 ppb for
9 annual mean due to the dominating role of ozone destruction by water vapor are found in four member
10 ensemble of CMIP6 ESM for surface warmings of 1.5 to 2.5 °C (Figure 6.14). This decrease is driven by the
11 ozone decrease over oceans, especially in the tropics (decrease of 1 ppb to 5 ppb) and large parts of the
12 continental unpolluted regions. The sensitivity of annual mean surface ozone to the level of surface warming
13 over these remote areas varies spatially from -0.2 to -2 ppb °C⁻¹ (see Figure 6.SM.1 in the Supplementary
14 Material)

15
16 Over ozone producing regions of the world, such as in North America, Europe, and East Asia, AR5 and post
17 AR5 model studies project a general increase of surface ozone levels (climate change penalty on ozone) in a
18 future warmer climate particularly during summertime (Fu and Tian, 2019). However, in current regional
19 models, using more robust protocols, this increase of surface ozone, attributable to climate change is of
20 lower magnitude than in previous estimates (Lacressonnière et al., 2016; Garcia-Menendez et al., 2017).
21 Climate change enhances the efficiency of precursor emissions to generate surface ozone in polluted regions,
22 (Schnell et al., 2016) and thus the magnitude of this effect will depend on the emissions considered in the
23 study (present or future, and mitigated or not) (Colette et al., 2015; Fiore et al., 2015).

24
25 Considering anthropogenic emissions of precursors globally higher than the current emissions (SSP3-7.0 in
26 2050, see Figure 6.20), the CMIP6 ensemble confirms the surface O₃ penalty due to climate change over
27 regions close to anthropogenic pollution sources or close to natural emission sources of ozone precursors
28 (e.g. biomass burning areas), with a penalty of a few ppb for the annual mean, proportional to warming
29 levels (Figure 6.14). This rate ranges regionally from 0.2 to 2 ppb °C⁻¹ (see Figure 6.SM.1 in Supplementary
30 Material). The CMIP6 ESMs show this consistently for South East Asia (in line with Hong et al., (2019) and
31 Schnell et al. (2016)) and for India (in line with (Pommier et al., 2018) as well as in parts of Africa and
32 South America, close to enhanced BVOC emissions (at least 3 out 4 ESMs agree on the sign of change). The
33 results are mixed in polluted regions of Europe and US because of lower anthropogenic precursor emissions
34 which leads to a very low sensitivity of surface ozone to climate change (-0.5 ppb °C⁻¹ to 0.5 ppb °C⁻¹) (see
35 Figure 6.SM.1 in Supplementary Material) and thus the ESMs can disagree on sign of changes for a given
36 warming level. This heterogeneity in the results is also found in regional studies over North America
37 (Gonzalez-Abraham et al., 2015; Val Martin et al., 2015; Schnell et al., 2016; He et al., 2018; Nolte et al.,
38 2018; Rieder et al., 2018) or over Europe (Colette et al., 2015; Lacressonnière et al., 2016; Schnell et al.,
39 2016; Fortems-Cheiney et al., 2017).

40
41 Overall, warmer climate is expected to reduce surface ozone in unpolluted regions as a result of greater water
42 vapor abundance accelerating ozone chemical loss (*high confidence*). Over regions with high anthropogenic
43 and/or natural ozone precursor emissions, there is prevailing evidence that climate change will introduce a
44 surface O₃ penalty increasing with increasing warming levels (with a magnitude ranging regionally from 0.2
45 to 2 ppb °C⁻¹) (*medium confidence to high confidence*). Yet, there are uncertainties in processes affected in a
46 warmer climate which can impact and modify future baseline and regional/local surface ozone levels. The
47 response of surface ozone to future climate change through stratosphere-troposphere exchange, soil NO_x
48 emissions and wildfires is positive (*medium confidence*). In addition, there is *low confidence* in the
49 magnitude of effect of climate change on surface ozone through biosphere interactions (natural CH₄, non-
50 methane BVOC emissions and ozone deposition) and lightning NO_x emissions.

1 [START FIGURE 6.14 HERE]

2
3 **Figure 6.14: Multi-model annual mean change in surface ozone (O₃) (ppb) concentrations at different warming**
4 **levels.** Changes are shown for a) 1.0°C, b) 1.5°C, c) 2.0°C and d) 2.5°C increase in global mean surface
5 air temperature. CMIP6 models include GFDL-ESM4, GISS-E2-1-G, MRI-ESM2-0 and UKESM1-0-LL.
6 For each model the change in surface O₃ is calculated as difference between the ssp370SST and
7 ssp370pdSST experiments in the year when the difference in the global mean surface air temperature
8 between the experiments exceeds the temperature threshold. The difference is calculated as a 20-year
9 mean in surface O₃ around the year when the temperature threshold in each model is exceeded. The multi-
10 model change in global annual mean surface O₃ concentrations with $\pm 1 \sigma$ are shown within parenthesis.
11 Uncertainty is represented using the simple approach: No overlay indicates regions with high model
12 agreement, 3 out of 4 models agree on sign of change; diagonal lines indicate regions with low model
13 agreement, where 3 out of 4 models agree on sign of change. For more information on the simple
14 approach, please refer to the Cross-Chapter Box Atlas.1. Further details on data sources and processing
15 are available in the chapter data table (Table 6.SM.1).

16
17 [END FIGURE 6.14 HERE]

18 19 20 6.5.2 Impact of climate change on particulate matter

21
22 Changes in concentration and chemistry of particulate matter (PM) in a changing climate depend in a
23 complex manner on the response of the multiple interactions of changes in emissions, chemical processes,
24 deposition and other factors (e.g., temperature, precipitation, circulation patterns). These changes are
25 difficult to assess and, at the time of AR5, *no confidence level* was attached to the overall impact of climate
26 change on PM_{2.5} (Kirtman et al., 2013). Possible changes induced by climate change may concern both
27 atmospheric concentration levels and chemical composition.

28
29 Higher temperatures increase the reaction rate of gaseous SO₂ to particulate sulphate conversion but also
30 favour evaporation of particulate ammonium nitrate (Megaritis et al., 2013). Also, higher temperatures are
31 expected to affect BVOC emissions (e.g. Pacifico et al., 2012) that would influence SOA concentrations,
32 although this effect has been questioned by more recent evidence (Wang et al., 2018a; Zhao et al., 2019c).
33 More generally, climate change will also affect dust concentration levels in the atmosphere (see Section
34 6.2.2.4) and the occurrence of forest fires, both very large sources of aerosols to the global troposphere (see
35 Section 6.2.1.2.3).

36
37 Wet deposition constitutes the main sink for atmospheric PM (Allen et al., 2016b, 2019b; Xu and Lamarque,
38 2018). In particular, precipitation frequency has a higher effect on PM wet deposition than precipitation
39 intensity (Hou et al., 2018). PM is also sensitive to wind speed and atmospheric stability conditions
40 emphasising the importance of stagnation episodes and low planetary boundary layer heights for increasing
41 PM atmospheric concentrations (Porter et al., 2015).

42
43 At the global scale, depending on its magnitude, the warming leads either to a small increase in global mean
44 PM concentration levels (ca. 0.21 $\mu\text{g m}^{-3}$ in 2100 for RCP8.5), mainly controlled by sulphate and organic
45 aerosols or a small decrease (-0.06 $\mu\text{g m}^{-3}$ for RCP2.6), Westervelt et al. (2016) and Xu and Lamarque
46 (2018). On the other hand, Xu and Lamarque (2018) and Allen et al. (2016c, 2019b) found an increase of
47 aerosol burden and PM surface concentration throughout the 21st Century, attributed to a decrease in wet
48 removal flux despite the overall projected increase in global precipitation, on the ground of an expected shift
49 of future precipitation towards more frequent heavy events. Based only on three models, the CMIP6
50 ensemble shows that for most land areas, there is low agreement between models on the sign of climate
51 change on annual mean PM_{2.5} (see Figure 6.SM.2 in Supplementary Material).

52
53 Due to the typical atmospheric lifetime of PM in the atmosphere, of the order of a few days, most studies
54 dealing with the future PM concentration levels have a regional character and concern mainly Europe
55 (Megaritis et al., 2013; Lacressonnière et al., 2016, 2017; Lemaire et al., 2016; Cholakian et al., 2019), the
56 US (Penrod et al., 2014; Fiore et al., 2015; Gonzalez-Abraham et al., 2015; Shen et al., 2017; He et al., 2018;

1 Nolte et al., 2018), South and East Asia (Jiang et al., 2013; Nguyen et al., 2019) and India (Pommier et al.,
2 2018). No studies are available for other areas of the world.

3
4 Changes in the chemical composition of PM as a result of future climate change can also be an important
5 issue for the effects of PM on human health and the environment, but only a few sparse data are available in
6 the literature on this and the results are, as yet, inconclusive (Im et al., 2012; Jiang et al., 2013; Megaritis et
7 al., 2013; Gonzalez-Abraham et al., 2015; Gao et al., 2018; He et al., 2018; Cholakian et al., 2019).

8
9 Overall, there is *medium confidence (medium evidence, high agreement)* in a small effect, positive or
10 negative, on PM global burden due to climate change.

11 12 13 **6.5.3 Impact of climate change on extreme pollution**

14
15 Extreme air pollution is identified as the concentration of an air pollutant that is above a given threshold
16 value (high concentration or a high percentile) as the sensitivity of peak values to meteorological conditions
17 can be different from sensitivity of the median or mean (Porter et al., 2015). AR5 assessed with medium
18 confidence that uniformly higher temperatures in polluted environments will trigger regional feedbacks in
19 chemistry and local emissions that will increase peak ozone and PM pollution, but assessed low confidence
20 in projecting changes in meteorological blocking associated with these extreme episodes.

21
22 Meteorological conditions, such as heat waves, temperature inversions, and atmospheric stagnation episodes
23 favor air quality extremes and are influenced by changing climate (Fiore et al., 2015). The body of literature
24 on the connection between climate change and extreme anthropogenic pollution episodes is essentially based
25 on correlation and regression applied to observation reanalysis but the metrics and methodologies differ
26 making quantitative comparisons difficult. Many emission processes in the natural systems are sensitive to
27 temperature, and burst of emissions as a response to extreme weather, as in the case of wildfires in dry
28 conditions (Bondur et al., 2020; Xie et al., 2020) can occur which would then add to the risk of extreme air
29 pollution but are not sufficiently constrained to be quantitatively assessed.

30
31 Since AR5, published studies provide augmented evidence for the connections between extreme ozone and
32 PM pollution events and high temperatures, especially long-lasting heat waves, whose frequency is
33 increasing due to a warming climate (Lelieveld et al., 2014; Porter et al., 2015; Hou and Wu, 2016; Schnell
34 and Prather, 2017; Zhang et al., 2017b) (Sun et al., 2017) Jing et al., 2017. However, relationship between air
35 pollution and individual meteorological parameters is exaggerated because of covariation on synoptic time
36 scales (Fiore et al., 2015). For example, heatwaves are often associated with clear-skies and stagnation,
37 making clear attribution to specific meteorological variable complicated. In Asia, future changes in winter
38 conditions have also been shown to favour more particulate pollution (Cai et al., 2017), and (Zou et al.
39 (2017). The relationship between the occurrence of stagnation episodes and high concentrations of ozone and
40 PM_{2.5} has been shown to be regionally and metric dependant (Oswald et al., 2015; Sun et al., 2017; Kerr and
41 Waugh, 2018; Schnell et al., 2018; Garrido-Perez et al., 2019).

42
43 The increase of heatwaves frequency, duration and intensity is extremely likely on all continents for different
44 future warming levels (Chapter 11, see Table 11.2 and Section 11.3.5). However, there is low confidence in
45 projected changes in storm tracks, jets, and blocking and thus their influence in extreme temperatures in mid-
46 latitudes (See Section 11.3.1).

47
48 In conclusion, there is still *medium confidence* that climate driven changes in meteorological conditions,
49 such as heatwaves or stagnations, will favour extreme air pollution episodes over highly polluted areas,
50 however the relationship between these meteorological conditions and high concentrations of ozone and
51 PM_{2.5} have been shown to be regionally and metric dependant.

6.6 Air Quality and Climate response to SLCF mitigation

LLGHG emission reductions are typically motivated by climate mitigation policies, whereas SLCF reductions mostly result from air pollution control, climate policies (see FAQ6.2) as well as policies focusing on achieving UN Sustainable Development Goals (SDGs) (see Box 6.2). The management of several SLCFs (BC, CH₄, tropospheric O₃, and HFCs) is considered in the literature as a fast-response, near-term measure to curb climate change, while reduction of emissions of LLGHGs is an essential measure for mitigating long-term climate warming (Shindell et al., 2012a, 2017a; Shoemaker et al., 2013; Rogelj et al., 2014b; Lelieveld et al., 2019). Note that the term short-lived climate pollutants (SLCPs) referring only to warming SLCFs has been used within the policy arena. The SR 1.5 report states that limiting warming to 1.5°C to achieve Paris Agreement goals, implies net zero CO₂ emissions around 2050 and concurrent deep reductions in emissions of non-CO₂ forcers, particularly methane (Rogelj et al., 2018a). In addition, several SLCFs are key air pollutants or precursors of fine particulate matter (PM_{2.5}) and tropospheric O₃ and therefore subject to control driven by air quality targets.

Policies addressing either SLCFs or LLGHGs reduction, often prioritize mitigation of specific anthropogenic sources, such as energy production, industry, transportation, agriculture, waste management, and residential fuel use. Choice of the targetted sector and chosen measures will determine the ratios of emitted SLCFs and LLGHGs. These changes in emissions of co-emitted species will result in diverse responses driven by complex chemical and physical processes and resulting climate perturbations (see Section 6.5). The understanding of the co-benefits through sectoral mitigation efforts (as well as potential negative impacts) is essential to inform policy making.

The discussion of targeted SLCF policies and their role in climate change mitigation ranges from critical evaluation of the climate co-benefits (Smith and Mizrahi, 2013; Pierrehumbert, 2014; Rogelj et al., 2014b; Strefler et al., 2014; Allen et al., 2016a), modelling the potential of dedicated BC and CH₄ reductions in association with or without climate policy (Harmsen et al., 2019; Smith et al., 2020), individual or multi-component mitigation in relation to natural variability (Samset et al., 2020), warning about the risk of diversion of resources from LLGHGs, especially CO₂, policies (e.g., Shoemaker et al., 2013), to seeing it as an opportunity to strengthen commitment and accelerate action on LLGHGs (Victor et al., 2015; Aakre et al., 2018).

Over the last decade, research on air quality-climate interactions and feedbacks has brought new attention of policy communities on the possibility of win-win mitigation policies that could both improve air quality and mitigate climate change, possibly also reducing the cost of interventions (Anenberg et al., 2012; Shindell et al., 2012a, 2017a, Schmale et al., 2014a, 2014b; Sadiq et al., 2017; Fay et al., 2018; Harmsen et al., 2020). Haines et al. (2017a) and Shindell et al. (2017a) connect the measures to mitigate SLCFs with the achievements of some of the SDGs. Indeed, most studies on co-benefits to date focus on the impacts of climate mitigation strategies, in particular to meet Nationally Determined Contributions (NDCs) and/or specific global temperature targets, on air quality and human health (West et al., 2013; Zhang et al., 2016; Rao et al., 2016; Shindell et al., 2016, 2017a; Chang et al., 2017; Haines et al., 2017a; Xie et al., 2018; Li et al., 2018a; Markandya et al., 2018; Williams et al., 2018; Lelieveld et al., 2019). Such co-benefits of climate mitigation for air quality and human health can offset the costs of the climate measures (Saari et al., 2015; Li et al., 2018a). A growing number of studies analyse the co-benefits of current and planned air quality policies on LLGHGs and global and regional climate change impacts (Lund et al., 2014a; Akimoto et al., 2015; Lee et al., 2016; Maione et al., 2016; Peng et al., 2017).

This section assesses the effects of mitigating SLCFs, motivated by various objectives, discussing temperature response time, temperature and air quality attribution of SLCFs sources, and chosen mitigation approach. The effects of the measures, to contain the spread of COVID-19 in 2020, on air quality (AQ) and climate are discussed in cross-chapter Box 6.1 at the end of this section.

6.6.1 Implications of lifetime on temperature response time horizon

The effect over time on GSAT following a mitigation effort affecting emissions of LLGHGs or SLCFs depends on the lifetimes of the LLGHGs and SLCFs, their radiative efficiencies, how fast the emissions are reduced, how long reductions last (limited time or sustained reduction), and the inertia of the climate system itself. Mitigation of SLCFs are often implemented through new legislation or technology standards for the different emission sectors and components, implying that reductions are sustained over time.

It is often perceived that the full climatic response following mitigation of SLCFs will occur almost immediately. However the inertia of the climate system strongly modifies the short-term and long-term response. SLCFs with lifetimes shorter than the time scales for interhemispheric mixing (1-2 years) can cause a more spatially heterogeneous forcing than LLGHGs and thus different regional patterns of the climate response (see Sections 6.4.3). The temporal response in GSAT to a radiative forcing can be quantified using linear impulse response functions (see Cross Chapter box 7.1), (Olivié and Peters, 2013; Geoffroy et al., 2013; Smith et al., 2018). Figure 6.15 shows the GSAT response for sustained step reduction in emissions of idealised SLCFs with different lifetimes. The response is relative to a baseline with constant emissions, so effects of emissions before the step reduction is not shown. For SLCFs with lifetimes shorter than a few years, the concentrations quickly reach a new steady state and the response time is primarily governed by the thermal inertia and thus the time scales of the climate system. For compounds with lifetime on the order of 10 years (for example methane), there is about a 10-year delay in the response during the first decades, compared to compounds with lifetimes less than one year. However on longer time scales the response is determined solely by the time scales of the climate system itself. For CO₂ (dashed line in Figure 6.15) the temporal response is very different due to the long time scale for mixing into the deep ocean and therefore a substantial fraction of atmospheric CO₂ is only removed on millenium time scales. This means that for SLCFs including methane, the rate of emissions drives the long term stabilisation, as opposed to CO₂ where the long term effect is controlled by cumulative emissions (Allen et al., 2018b). Methods to compare rates of SLCF emission with cumulative CO₂ emissions are discussed in chapter 7, section 7.6.1.4.

[START FIGURE 6.15 HERE]

Figure 6.15: Global mean surface air temperature (GSAT) response to an abrupt reduction in emissions (at time $t=0$) of idealized climate forcing agents with different lifetimes. All emissions are cut to give a radiative forcing of -1 W m^{-2} at steady state (except for CO₂). In other words, if the yearly emissions are E_0 before the reduction, they will have a fixed lower value $E_{\text{year}>0} = (E_0 - \Delta E)$ for all succeeding years. For comparison, the GSAT response to a sustained reduction in CO₂ emissions resulting in an RF of -1 W m^{-2} in year 100 is included (dashed line). The temperature response is calculated using an impulse response function (see Cross-Chapter Box 7.1) with a climate feedback parameter of $-1.31 \text{ W m}^{-2} \text{ }^\circ\text{C}^{-1}$. Further details on data sources and processing are available in the chapter data table (Table 6.SM.1).

[END FIGURE 6.15 HERE]

As a consequence, in idealized ESM studies that assume an instantaneous removal of all anthropogenic or fossil fuel-related emissions, a rapid change in aerosol levels occurs leading to large increases in GSAT with the rate of warming lasting for several years. Similarly, the thermal inertia causes the pulse emissions (Figure 6.16) of SLCFs to have a significant effect on surface temperature even after 10 years.

In summary, for SLCFs with short lifetime (e.g. months), the response in surface temperature occurs strongly as soon as a sustained change in emissions is implemented and continues to grow for a few years, primarily due to thermal inertia in the climate system (*high confidence*). Near its maximum, the response slows down but will then take centuries to reach equilibrium (*high confidence*). For SLCF with longer lifetimes (e.g. a decade), a delay equivalent to their lifetimes comes in addition to the delay due the thermal inertia (*high confidence*).

6.6.2 Attribution of temperature and air pollution changes to emission sectors and regions

Assessment of the temperature response to source emission sectors is important for identifying priority mitigation measures and designing efficient mitigation strategies.

Temperature effects of emissions can be quantified for the historical contribution to the present temperature impact (see 6.4.2), for idealized one-year pulses of emissions or for continued (sustained) emissions at present levels and for changes during a specific time period, for emissions from future scenarios with various hypotheses, giving complementary information to feed mitigation strategies. AR5 assessed the net global temperature impact of source emission sectors from a one-year pulse (single year worth) of year 2008 emissions and found that the largest contributors to warming on 50-100 year time scales are the energy, industrial and on-road transportation sectors. Sectors that emit large amounts of methane (agriculture and waste management) and black carbon (residential biofuel) are important contributors to warming over short time horizons up to 20 years. Below, we discuss the effect on ERFs, temperature and air pollution of selected key sectors estimated to have large non-CO₂ forcing, including agriculture, residential and commercial, and transport (aviation, shipping, land transportation).

6.6.2.1 Agriculture

According to the SRCCL assessment (Jia et al., 2019), agriculture, forestry and other land use (AFOLU) is a significant net source of GHG emissions (*high confidence*), with more than half of these emissions attributed to non-CO₂ GHGs from agriculture. With respect to SLCFs, agricultural activities are major global sources of CH₄ and NH₃ (Section 6.2.1). The agriculture sector exerts strong near term warming due to large CH₄ emission that is slightly offset by a small cooling from secondary inorganic aerosols formed notably from the NH₃ emission (Heald and Geddes, 2016b; Lund et al., 2020). For present day emissions, agriculture is the 2nd largest contributor to warming on short time scales but with a small persisting effect on surface temperature (+0.0012±0.00028°C) after a pulse of current emissions (Lund et al., 2020 and Figure 6.16, see detailed description in 6.6.2.3.4). Aerosols produced from agricultural emissions, released after nitrogen fertilizer application and from animal husbandry, influence surface air quality and make an important contribution to surface PM_{2.5} in many densely populated areas (Lelieveld et al., 2015b; Bauer et al., 2016) and Figure 6.17.

6.6.2.2 Residential and Commercial cooking, heating

The residential and commercial sector is associated with SLCF emissions of carbonaceous aerosol, CO and VOCs, SO₂ and NO_x and can be split by fuel type (biofuel or fossil fuel) where residential fossil fuel is also associated with CO₂ and CH₄ emissions (Section 6.2.1). The impacts of residential CO and VOC emissions are warming and SO₂ and NO_x are net cooling. However, the net sign of the global radiative effects of carbonaceous aerosols from the residential sector and solid fuel cookstove emissions (warming or cooling) is not well constrained based on evidence from recent global atmospheric modelling studies. Estimates of global residential sector direct aerosol-radiation effects range from -20 to +60 mW m⁻² (Kodros et al., 2015) and -66 to +21 mW m⁻² (Butt et al., 2016); and aerosol-cloud effects range from -20 to +10 mWm⁻² (Kodros et al., 2015) and -52 to -16 mW m⁻² (Butt et al., 2016). Uncertainties are due to assumptions about the aerosol emission masses, size distribution, aerosol optical properties and mixing states (see Section 6.3.5.3). Allowing BC to act as an INP in a global model leads to a much larger global forcing estimate from -275 to +154 mW m⁻² with large uncertainty range due to uncertainty in the plausible range of maximum freezing efficiency of BC (Huang et al., 2018). The net sign of the impacts of carbonaceous aerosols from residential burning on radiative forcing and climate (warming or cooling) is ambiguous. The residential biofuel sector is a major concern for indoor air quality (Bonjour et al., 2013). In addition, several atmospheric modelling studies find that this sector is also important for outdoor air quality and even a dominant source of population-weighted outdoor PM_{2.5} in India and China (Lelieveld et al., 2015b; Silva et al., 2016a; Spracklen et al., 2018, Figure 6.17; Reddington et al., 2019).

1 The net climate impact of the residential sector is warming in the near term of $+0.0018\pm 0.00084^{\circ}\text{C}$ for fossil
2 fuel use and $+0.0014\pm 0.0012^{\circ}\text{C}$ for biofuel use, and $+0.0017\pm 0.00017^{\circ}\text{C}$ and $+0.0001\pm 0.000079^{\circ}\text{C}$,
3 respectively over a 100 years horizon, after a one year pulse of current emissions (Lund et al., 2020), due to
4 the effects of BC, CH₄, CO and VOCs which adds to that of CO₂ but the uncertainty in the net sign of
5 carbonaceous aerosol impacts challenges overall quantitative understanding of this sector and leads to *low*
6 *confidence* in this assessment. Residential sector emissions are an important source of indoor and outdoor air
7 pollution in Asia and globally (*high confidence*).
8
9

10 6.6.2.3 Transportation

11 6.6.2.3.1 Aviation

12 Aviation is associated with a range of SLCFs from emissions of NO_x, aerosol particles, and NO_x alongside
13 its emissions of water vapour and CO₂. The largest SLCF effects are those from the formation of persistent
14 condensation trails (contrails) and NO_x emissions. Persistent contrails are ice crystal clouds, formed around
15 aircraft soot particles (and water vapour from the engine), when the ambient cold and ice-supersaturated
16 atmosphere, which can spread and form contrail cirrus clouds. The ‘net NO_x’ effect arises from the formation
17 of tropospheric O₃, counterbalanced by the destruction of ambient CH₄ and associated cooling effects of
18 reductions in stratospheric water vapour and background O₃. AR5 assessed the radiative forcing from
19 persistent linear contrails to be $+0.01$ ($+0.005$ to $+0.03$) W m⁻² for year 2011, with a medium confidence
20 (Boucher et al., 2013). The combined linear contrail and their subsequent evolution to contrail cirrus
21 radiative forcing from aviation was assessed to be $+0.05$ ($+0.02$ to $+0.15$) W m⁻², with a low confidence. An
22 additional forcing of $+0.003$ W m⁻² due to emissions of water vapour in the stratosphere by aviation was also
23 reported (Boucher et al., 2013). The aviation sector was also estimated to lead to a net surface warming at 20
24 and 100 years horizons following a one year pulse emission. This net temperature response was determined
25 by similar contributions from contrails and contrail cirrus and CO₂ over a 20 year time horizon, and
26 dominated by CO₂ in a 100 years perspective (Figure 8.34 AR5: Myhre et al., 2013).
27
28

29 Our assessment builds on Lee et al. (2020a). Their study consists of an updated, comprehensive assessment
30 of aviation climate forcing in terms of RF and ERF based on a large number of studies and the most recent
31 air traffic and fuel use datasets available (for 2018), new calculations and the normalization of values from
32 published modeling studies, and combining the resulting best estimates via a Monte-Carlo analysis. Lee et al.
33 (2020a), reports a net aviation ERF for year 2018 emissions of $+0.101$ W m⁻² (5–95% likelihood range of
34 0.055 – 0.145) with major contributions from contrail cirrus (0.057 W m⁻²), CO₂ (0.034 W m⁻²), and NO_x
35 (0.017 W m⁻²). Contrails and aviation-induced cirrus yields the largest individual positive ERF term, whose
36 confidence level is similarly assessed to Lee et al. (2020a), as ‘low’ by Chapter 7 (Section 7.3.4.2) due to
37 potential missing processes, followed by CO₂ and NO_x emissions (Lee et al., 2020a). The formation and
38 emission of sulphate aerosol yields a negative (cooling) term. SLCF forcing terms contribute about 8 times
39 more than CO₂ to the uncertainty in the aviation net ERF in 2018 (Lee et al., 2020a). The largest uncertainty
40 in assessing aviation climate effects is on the interactions of BC and sulphate aerosols on cirrus and mixed
41 phase clouds, for which no best estimates of the ERFs were provided (Lee et al., 2020a).
42

43 One of the most significant changes between AR5 and AR6 in terms of aviation SLCFs is the explicit
44 calculation of a contrail cirrus ERF found to be 35% of the corresponding RF Bickel et al. (2020), which has
45 confirmed studies indicating smaller efficacy of linear contrails Ponater et al. (2005) and Rap et al. (2010).
46 The net-NO_x term is generally agreed to be a positive RF in the present day, although attribution in a non-
47 linear chemical system is problematic (Grewe et al., 2019) but Skowron et al. (2021) point out that the sign
48 of net NO_x term is dependent on background conditions and could be negative under certain future scenarios.
49

50 The best estimate ERFs from aviation (Lee et al., 2020b) have been used to calculate aviation-specific
51 Absolute Global Temperature change Potential (AGTP) using the method described in (Lund et al., 2020)
52 and subsequently compute the effect of one-year pulse of aviation emissions on global-mean surface
53 temperature on a 10- and 100-year time horizon (Figure 6.16, see 6.6.2.3.4). The effect of contrail-cirrus is
54 most important for the estimated net GSAT response after the first decade, followed by similar warming
55 contributions from NO_x and CO₂ emissions. At a 20-year time horizon, the net contribution from aviation to

1 GSAT has switched from positive to a small negative effect (see 6.SM.4). This is due to the combination of
2 rapidly decaying contrail-cirrus warming and the complex time variation of the net temperature response to
3 NO_x emissions, which changes sign between 10 and 20 years due to the balance between the positive short-
4 lived ozone forcing and negative forcing from changes in CH₄ and CH₄-induced changes in ozone and
5 stratospheric water vapour. The net GSAT response to aviation emissions has previously been estimated to
6 be positive on a 20-year time horizon (Chapter 8 AR5; Lund et al., 2017). This difference in net GSAT after
7 20 years in AR5 compared to AR6 results primarily from a shorter time scale of the climate response in the
8 underlying AGTP calculations in Lund et al. (2020), which means the initial, strong impacts of the most
9 short-lived SLFCs, including the warming by contrail-cirrus decay faster, in turn giving the net NO_x effect a
10 relatively higher importance after 20 years. On longer time horizons, the net GSAT response switches back
11 to positive, as CO₂ becomes the dominating warming contribution.

12
13 In summary, the net aviation ERF is assessed to be +0.1 W m⁻² (±0.045) for the year 2018 (*low confidence*).
14 This confidence level is largely a result of the fact that the SLCF-related terms which counts for more than
15 half (66% of the net aviation ERF) are the most uncertain terms. The climate response to SLCF-related
16 aviation terms exhibits substantial spatiotemporal heterogeneity in characteristics (*high confidence*). Overall,
17 cirrus and contrail cirrus warming, as well as NO_x-induced ozone increase, induce strong, but short-lived
18 warming contributions to the GSAT response ten years after a one-year pulse of present-day aviation
19 emissions (*medium confidence*), while CO₂ both gives a warming effect in the near term and dominates the
20 long-term warming impact (*high-confidence*).

21 22 23 6.6.2.3.2 Shipping

24 Quantifying the effects of shipping on climate is particularly challenging because (i) the sulphate cooling
25 impact is dominated by aerosol cloud interactions and (ii) ship emissions contain NO_x, SO_x and BC, which
26 lead to mixed particles. Previous estimates of the sulphate radiative effects from present day shipping span
27 the range -47 to -8 mW m⁻² (direct radiative effect) and -600 to -38 mW m⁻² (indirect radiative effects)
28 (Balkanski et al., 2010; Eyring et al., 2010; Lauer et al., 2007; Lund et al., 2012). Partanen et al. (2013)
29 reported a global mean ERF for year 2010 shipping aerosol emissions of -390 mW m⁻². The temperature
30 change has been shown to be highly sensitive to the choice of aerosol-cloud parameterization (Lund et al.,
31 2012). One year of global present-day shipping emissions, not considering impact of recent low sulphur fuel
32 regulation (IMO, 2016), are estimated to cause net cooling in the near term (-0.0024±0.0025°C) and slight
33 warming (+0.00033±0.00015°C) on a 100-year horizon (Lund et al., 2020).

34
35 Shipping is also of importance for air pollution in coastal areas along the major trade routes, especially in
36 Europe and Asia (Corbett et al., 2007; Liu et al., 2016b, Figure 6.17; Jonson et al., 2020). Jonson et al.
37 (2020) estimated that shipping is responsible for 10% or more of the controllable PM_{2.5} concentrations and
38 depositions of oxidised nitrogen and sulphur for many coastal countries. Widespread introduction of low-
39 sulphur fuels in shipping from 2020 (IMO, 2016) will lead to improved air quality and reduction in
40 premature mortality and morbidity (Sofiev et al., 2018).

41
42 In summary, one year of present-day global shipping emissions (i.e., without the implementation of the 2020
43 clean fuel standards) cause a net global cooling (-0.0024±0.0025°C) on 10-20 year time horizons (*high*
44 *confidence*) but its magnitude is of *low confidence*.

45 46 47 6.6.2.3.3 Land transportation

48 The on-road and off-road transportation sectors have a net warming impact on climate over all time scales
49 (Berntsen and Fuglestvedt, 2008; Fuglestvedt et al., 2008; Unger et al., 2010; Lund et al., 2020). One year
50 pulse of present day emissions has a small net global temperature effect on short time-scales
51 (+0.0011±0.0045°C), predominantly driven by CO₂ and BC warming offset by NO_x-induced cooling through
52 CH₄ lifetime reductions (Lund et al., 2020).

53
54 The vehicle tailpipe emission profiles of diesel and gasoline are distinctly different. Diesel air pollutant
55 emissions are dominated by BC and NO_x whereas petrol air pollutant emissions are dominated by CO and

1 NMVOCs, especially when distribution and upstream losses are considered. Thus, net radiative effect of the
2 on-road vehicle fleets depends upon the share of different fuels used, in particular gasoline and diesel (Lund
3 et al., 2014b; Huang et al., 2020b). The net SLCF for year 2010 emissions from the global diesel vehicle
4 fleet have been estimated to be +28 mW m⁻² (Lund et al., 2014b). Huang et al. (2020b) estimated net global
5 radiative effects of SLCFs (including aerosols, O₃, and CH₄) from the gasoline and diesel vehicle fleets in
6 year 2015 at +13.6 and +9.4 mW m⁻², respectively, with similar fractional contributions of SLCFs to the total
7 global climate impact including CO₂ on the 20-year time scale (14-15%).

8
9 There is a consensus that on-road transportation sector emissions, including gasoline and diesel, are
10 important anthropogenic contributors to elevated surface O₃ and PM_{2.5} concentrations (Chambliss et al.,
11 2014; Lelieveld et al., 2015b; Silva et al., 2016a, Figure 6.17; Anenberg et al., 2019). At a global-scale, land
12 transportation has been estimated to be the dominant contributor to surface ozone concentrations in
13 populated areas (Silva et al., 2016b) and O₃-induced vegetation damages (Section 6.4.4) (Unger et al., 2020).
14 Furthermore, it is now well established that real world diesel NO_x emission rates are substantially higher, the
15 so-called “excess NO_x”, in all regional markets than in laboratory tests worsening air quality (Anenberg et
16 al., 2017; Jonson et al., 2017; Chossière et al., 2018) and contributing to slightly larger warming on the scale
17 of years and smaller warming at the decadal scale (Tanaka et al., 2018). Excess NO_x emissions from key
18 global diesel markets are estimated at 4.6 Tg a⁻¹ in 2015, with annual mean O₃ and PM_{2.5} increases of 1 ppb
19 and 1 μg m⁻³ across large-regions of Europe, India and China (Anenberg et al., 2017).

20
21 In summary, the present-day global land-based transport pulse emissions cause a net global warming on all
22 time scales (*high confidence*) and are detrimental to air quality (*high confidence*).

23 24 25 6.6.2.3.4 GSAT response to emission pulse of current emissions

26 Figure 6.16 presents the GSAT response to an idealized pulse of year 2014 emissions of individual SLCF
27 and LLGHG. The GSAT response is calculated for 11 sectors and 10 regions accounting for best available
28 knowledge and geographical dependence of the forcing efficacy of different SLCFs (Lund et al., 2020). Two
29 time horizons are shown (10 and 100 year) to represent near- and long-term effects (and 20 year horizon is
30 presented in Figure 6.SM.3). Other time horizon choices may affect the relative importance, and even sign in
31 the case of the NO_x effect, of the temperature response from some of the SLCFs, or be more relevant for
32 certain applications. GSAT response is calculated using the concept of AGTP (see Section 7.6.2.2). Further
33 details of the AGTP emulator applied in Figure 6.16 are provided in (Lund et al., 2020) and 6.SM.4 (see also
34 Section 7.6.1.2, cross-chapter Box 7.1 and 7.SM.7.2). As discussed in Lund et al. (2020), the AGTP
35 framework is primarily designed to study relative importance of individual emissions and sources, but the
36 absolute magnitude of temperature responses should be interpreted with care due to the linearity of the
37 AGTP, which does not necessarily capture all the non-linear effects of SLCFs emissions on temperature.

38
39 Differences in the mix of emissions result in net effects on GSAT that vary substantially, in both magnitude
40 and sign, between sectors and regions. SLCFs contribute substantially to the GSAT effects of sectors on
41 short time horizons (10-20 years) but CO₂ dominates on longer time horizons (Figure 6.16). As the effect of
42 the SLCFs decays rapidly over the first few decades after emission, the net long-term temperature effect is
43 predominantly determined by CO₂. N₂O adds a small contribution to the long-term effect of agriculture. CO₂
44 emissions cause an important contribution to near-term warming that is not always fully acknowledged in
45 discussions of LLGHGs and SLCFs (Lund et al., 2020).

46
47 The global sectoral ranking for near- and long-term global temperature effects is similar to the AR5
48 assessment despite regional reductions in aerosol precursor emissions between 2008 and 2014. This report
49 has applied updated climate policy metrics for SLCFs and treatment of aerosol-cloud interactions for SO₂,
50 BC and OC (Lund et al., 2020). By far the largest 10-year GSAT effects are from the energy production
51 (fossil fuel mining and distribution), agriculture and waste management (*high confidence*). CH₄ is the
52 dominant contributor in the energy production, agriculture, and waste management sectors. On the 10-year
53 time horizon, other net warming sectors are residential fossil fuel and energy combustion (dominated by
54 CO₂) and aviation and residential biofuel (dominated by SLCFs and cloud) (*medium confidence*). The total
55 residential and commercial sector including biofuel and fossil fuels is the 4th most warming economic sector

1 globally on short time horizons of 10-20 years. The energy combustion sector has considerable cooling from
2 high emissions of SO₂ that result in a relatively small net GSAT temperature effect on short time horizons,
3 despite the high CO₂ emissions from this activity. On the 10-year time horizon, global emissions from
4 industry and shipping cause a net cooling effect despite a considerable warming from CO₂ emissions. On the
5 100-year time horizon, the net effects of agriculture and waste management are small, while energy
6 combustion is the largest individual contributor to warming due to its high CO₂ emissions. The second
7 largest driver of long-term temperature change is industry, demonstrating the importance of non-CO₂
8 emissions for shaping relative weight over different time frames. Transport contributes a small net warming
9 on the 10-year time horizon that increases by a factor of three on the 100-year time horizon. In contrast,
10 aviation sector contribution to warming shrinks by about a factor of three between the 10- and 100-year time
11 horizons. Results for the 20-year time horizon are provided in the Supplementary Material 6.SM.4.
12 Compared to the 10-year time horizon, there are some changes in ranking, especially of sectors and regions
13 with a strong SO₂ contribution, which decays substantially between 10 and 20 years. Aviation is the sector
14 with most distinct difference between 10 and 20 year time horizons such that the net GSAT effect after 20
15 years becomes small but negative due to a switch in sign for the NO_x AGTP for this sector and these two
16 time horizons related to the stronger short-lived ozone response than other sectors.
17

18 In terms of source regions, the largest contributions to net short-term warming are caused by emissions in
19 Eastern Asia, Latin America and North America, followed by Africa, Eastern Europe and West-Central Asia
20 and South East Asia (*medium confidence*). However, the relative contributions from individual species vary.
21 In East Asia, North America, Europe and South Asia the effect of current emissions of cooling and warming
22 SLCFs approximately balance in the near term and these regions cause comparable net warming effects on
23 10- and 100-year time horizons (Figure 6.16). In Latin America, Africa and South East Asia and Developing
24 Pacific, CH₄ and BC emissions are currently high while emissions of CO₂ and cooling aerosols are low
25 compared to other regions, resulting in a net warming effect after 10 years that is substantially higher than
26 that of CO₂ alone.
27

28 Overall, the global sectors that contribute the largest warming on short time scales are the CH₄-dominated
29 sources, i.e., energy production (fossil fuel mining and distribution), agriculture and waste management
30 (*high confidence*). On short time scales, other net warming sectors are residential fossil fuel and energy
31 combustion (dominated by CO₂) and aviation and residential biofuel (dominated by SLCFs) (*medium*
32 *confidence*). On short-time scales, global emissions from industry and shipping cause a net cooling effect
33 despite a considerable warming from CO₂ emissions (*high confidence*). On longer time horizons, the sectors
34 that contribute the largest warming are energy combustion and industry due to the large CO₂ emissions (*high*
35 *confidence*).
36
37

38 6.6.2.3.5 Source attribution of regional air pollution

39 The attribution of present-day surface PM_{2.5} and O₃ concentrations to sectors and regions (Figure 6.17) is
40 based on 2014 CMIP6 emissions used in the TM5-FASST model (Van Dingenen et al., 2018) that has been
41 widely applied to analyse air quality in regional and global scenarios (Van Dingenen et al., 2009; Rao et al.,
42 2016, 2017; Vandyck et al., 2018; e.g., Harmsen et al., 2020). Regions with largest year 2014 population-
43 weighted annual average surface PM_{2.5} concentrations are South Asia, East Asia, and Middle East. The
44 dominant anthropogenic source of ambient PM_{2.5} in South Asia is residential and commercial sector
45 (biomass and coal fuel-based cooking and heating) with secondary contributions from energy and industry.
46 In East Asia, the main anthropogenic sources of ambient PM_{2.5} are energy, industry and residential sources.
47 Natural sources, predominantly dust, are the most important PM_{2.5} source in the Middle East, Africa and
48 Eurasia, contributing about 40-70% of ambient annual average concentrations (Figure 6.17). Agriculture is
49 an important contributor to ambient PM_{2.5} in Europe and North America, while open biomass burning is a
50 major contributor in South East Asia and Developing Pacific, North America as well as Latin America.
51 These results are consistent with several global and regional studies, where contribution of emission sources
52 to ambient PM_{2.5} or premature mortality was estimated at different scales (Guttikunda et al., 2014; Lelieveld
53 et al., 2015b; e.g., Amann et al., 2017; Qiao et al., 2018; Venkataraman et al., 2018; Wu et al., 2018).
54

55 Natural sources contribute more than 50% to surface ozone in all regions except South Asia and South East

1 Asia. South Asia, East Asia and the Middle East experience the highest surface ozone levels of all regions.
 2 For ozone, the anthropogenic sectoral attribution is more uniform across regions than for PM_{2.5}, except for
 3 South and Southeast Asia where land transportation plays a larger role and East Asia with most significant
 4 contribution from energy and industry. Land transportation and energy are the most important contributors to
 5 ozone across many of the regions, with smaller contributions from agriculture, biomass burning, waste
 6 management and industry. Open biomass burning is not a major contributor to surface ozone, except for
 7 Africa, Latin America and South East Asia where its contribution is estimated at about 5-10% of
 8 anthropogenic sources. Relative importance of natural and anthropogenic emission sources on surface ozone
 9 have been assessed in several studies (Uherek et al., 2010; Zare et al., 2014; Mertens et al., 2020; Unger et
 10 al., 2020) and the results are comparable with the estimates of the TM5-FASST used here.

11
 12 Residential and commercial cooking and heating is among the most important anthropogenic sources of
 13 ambient PM_{2.5}, except Middle East and Asia-Pacific Developed (*high confidence*) and agriculture is the
 14 dominant source in Europe and North America (*medium confidence*). Energy and industry are important
 15 PM_{2.5} contributors in most regions, except Africa (*high confidence*). Energy and land transportation are the
 16 major anthropogenic sources of ozone across many world regions (*medium to high confidence*).

17
 18
 19 [START FIGURE 6.16 HERE]

20
 21 **Figure 6.16: Global-mean surface temperature response 10 and 100 years following one year of present-day**
 22 **(year 2014) emissions.** The temperature response is broken down by individual species and shown for
 23 total anthropogenic emissions (top), sectoral emissions (left) and regional emissions (right). Sectors and
 24 regions are sorted by (high-to-low) net temperature effect on the 10-year time scale. Error bars in the top
 25 panel show uncertainty (5-95% interval) in net temperature effect due to uncertainty in radiative forcing
 26 only (calculated using a Monte Carlo approach and best estimate uncertainties from the literature - see
 27 Lund et al. (Lund et al., 2020) for details). CO₂ emissions are excluded from open biomass burning and
 28 residential biofuel use due to their unavailability in CEDS and uncertainties around non-sustainable
 29 emission fraction. Emissions for 2014 originate from the Community Emissions Data System (CEDS)
 30 (Hoesly et al., 2018), except for HFCs which are from Purohit et al. (2020), open biomass burning from
 31 van Marle et al. (2017), and aviation H₂O which is from Lee et al. (2020b). The split of fossil fuel
 32 production and distribution and combustion for energy and residential and commercial fuel use into fossil
 33 fuel and biofuel components obtained from the GAINS model (ECLIPSE version 6b dataset). Open
 34 biomass burning emissions are not included for the regions. Emissions are aggregated into fossil fuel
 35 production and distribution (coal mining, oil and gas production, upstream gas flaring, gas distribution
 36 networks), agriculture (livestock and crop production), fossil fuel combustion for energy (power plants),
 37 industry (combustion and production processes, solvent use losses from production and end use),
 38 residential and commercial (fossil fuel use for cooking and heating as well as HFCs leakage from A/C and
 39 refrigeration), waste management (solid waste, incl. landfills and open trash burning, residential and
 40 industrial waste water), land transportation (road and off-road vehicles, and HFC leakage from A/C and
 41 refrigeration equipment), residential and commercial (biofuels use for cooking and heating), open
 42 biomass burning (forest, grassland, savannah fires, and agricultural waste burning), shipping (incl.
 43 international shipping), and aviation (incl. international aviation). Further details on data sources and
 44 processing are available in the chapter data table (Table 6.SM.1).

45
 46 [END FIGURE 6.16 HERE]

47
 48
 49 [START FIGURE 6.17 HERE]

50
 51 **Figure 6.17: Emission source-sector attribution of regional population weighted mean concentrations of PM_{2.5}**
 52 **and ozone for present day emissions (year 2014).** Regional concentrations and source apportionment
 53 calculated with the TM5-FASST model (Van Dingenen et al., 2018) for the 2014 emission data from the
 54 Community Emissions Data System (CEDS) (Hoesly et al., 2018) and van Marle et al. (2017) for open
 55 biomass burning. PM_{2.5} dust and seasalt are monthly mean climatological average over 2010-2018 from
 56 CAMS global reanalysis (EAC4) (Inness et al., 2019), generated using Copernicus Climate Change
 57 Service information [January 2020]. Anthropogenic sectors are similar to those in Figures 6.3 and 6.16
 58 except grouping of fossil fuel production, distribution and combustion for energy under “Energy” and

1 grouping of use of fossil fuel and biofuel use for cooking and heating under “Residential and
 2 Commercial”. Further details on data sources and processing are available in the chapter data table (Table
 3 6.SM.1).
 4

5 **[END FIGURE 6.17 HERE]**
 6

8 **6.6.3 Past and current SLCF reduction policies and future mitigation opportunities**

9
 10 Several SLCF emission reduction strategies have been explored in the literature or are already pursued as
 11 part of the environmental and development policies, including air quality, waste management, energy
 12 poverty, and climate change. The effects of various policies and strategies have been addressed in a limited
 13 number of modelling studies having different objectives and range from assessment of specific policies and
 14 their regional effects (UNEP and WMO, 2011; Shindell et al., 2012a, 2017a, AMAP, 2015a, 2015b; Haines
 15 et al., 2017a; UNEP and CCAC, 2018; Harmsen et al., 2019; UNEP, 2019) to large-scale global scenario
 16 studies with varying level of SLCF control (Sand et al., 2016; e.g., Rogelj et al., 2018b; Shindell and Smith,
 17 2019). They could be grouped into:
 18

- 19 – Projections of future SLCF emissions compatible with the climate mitigation trajectories
 20 investigated in the climate model intercomparison projects (respective RCP or SSP scenarios): In
 21 such scenarios, when climate change mitigation is considered, it is associated with strong decrease of
 22 CO₂ emissions, largely relying on fossil fuel use reduction, the co-emitted SLCFs from combustion
 23 and methane from production and distribution of fossil fuels will be reduced proportionally.
 24 Depending on carbon price and climate mitigation target, further reduction of methane from waste
 25 and agriculture will also be part of such scenarios. Limitations of RCP scenarios, where continuous
 26 strengthening of air quality legislation was assumed and therefore lack of futures where global and
 27 regional air quality deteriorates, for the analysis of air quality and potential for mitigation of SLCFs
 28 have been discussed in literature (e.g., Amann et al., 2013; Von Schneidemesser et al., 2015). SSP
 29 scenarios consider various levels of air pollution control, in accordance with their socioeconomic
 30 narrative, and thus cover a wider span of SLCF trajectories (Section 6.7). The economical cost of
 31 implementation of these scenarios and their co-benefits on air quality and SDGs are assessed in
 32 WGIII (Chapter 3) reports.
 33
- 34 – Projections of SLCF emissions assuming strong reduction of all air pollutants in the absence of
 35 climate mitigation (e.g., the SSP3-LowSLCF scenario); the latter is an idealized simulation of a very
 36 ambitious air quality policy where maximum technical potential of existing end-of-pipe technologies
 37 is explored in the SSP3-7.0 scenario. Methane reduction can also be part of such sensitivity analysis
 38 although methane reductions have not historically been motivated by air pollution concerns.
 39
- 40 – Projection of emissions targeting air quality or other development priorities: Anthropogenic
 41 emissions source-structure and the level of exposure to pollution and subsequent effects vary
 42 significantly from one region to another (section 6.3.2). Therefore, the air quality policy, regional
 43 climate impact concerns as well as development priorities, and consequently the level of mitigation
 44 of particular SLCF species, will differ regionally and source-wise with respect to the emissions
 45 sources, influence of intercontinental transport of pollution, and spatial physical heterogeneities
 46 (Lund et al., 2014b; AMAP, 2015a; Sand et al., 2016; Turnock et al., 2016; Sofiev et al., 2018;
 47 WMO, 2018). This is also the case at a finer local and regional scale where priorities and scope for
 48 SLCF mitigation will differ (e.g., Amann et al., 2017; UNEP, 2019).
 49
- 50 – Projections exploring mitigation potential for a particular source or SLCF: These studies focus on
 51 the assessment of SLCF reduction potential that can be realised with either existing and proven
 52 technologies or extend the scope to include transformational changes needed to achieve further
 53 reduction (UNEP and WMO, 2011; Stohl et al., 2015; Velders et al., 2015; Purohit and Höglund-
 54 Isaksson, 2017; e.g., Gómez-Sanabria et al., 2018; UNEP, 2019; Höglund-Isaksson et al., 2020;
 55 Purohit et al., 2020) – several of these studies are subsequently used for parameterization of the

1 models used to develop emission scenarios (e.g., IAM models used in the IPCC process).

2
3 In the following subsections, we assess the SLCF mitigation and its effects as identified in regional and
4 global studies evaluating past and current air quality and other SLCF regulations (Sections 6.6.3.1, 6.6.3.2,
5 6.6.3.3). Development policies, independent from the CMIP6 assessment framework, including peer-
6 reviewed studies and initiatives like UNEP, analysing win-win climate, air quality and SDG motivated
7 strategies are discussed in Section 6.6.3.4. Note that sensitivity studies where impacts of complete removal
8 of particular species are analysed (e.g., Samset et al., 2018) are used sparingly in this assessment. While such
9 analysis can be useful for assessing the effect of a zero-emission commitment (see Chapter 4.7.1.1), they do
10 not correspond to a realistic SLCF mitigation strategy with plausible pace of implementation and removal of
11 co-emitted species (Shindell and Smith, 2019). Discussion of climate and air quality implications of SLCF
12 reductions in SSP scenarios is provided in section 6.7.

13 14 15 *6.6.3.1 Climate response to past AQ policies*

16
17 Air quality policies emerged several decades ago focusing on emission mitigation first driven by local then
18 by regional scale air quality and ecosystem damage concerns, that is, health impacts and acidification and
19 eutrophication. They have made it possible to reduce or limit pollution exposure in many megacities or
20 highly populated regions e.g. in Los Angeles, Mexico City, Houston in North America (Parrish et al., 2011),
21 Santiago in Chile (Gallardo et al., 2018), São Paulo in Brazil (Andrade et al., 2017), Europe (Reis et al.,
22 2012; Crippa et al., 2016; Serrano et al., 2019), and over East Asia during the last decade (Silver et al., 2018;
23 Zheng et al., 2018b). However, very few studies have quantified the impact of these policies on climate. AR5
24 concluded that air quality control will have consequences on climate including strong regional variability,
25 however, no estimates of impacts of specific air quality policy were available. Since AR5, few studies have
26 provided estimates of climate relevant indicators affected by significant air pollutant burden changes due to
27 air quality policy in selected regions. Turnock et al. (2016) estimated that the strong decrease in NO_x, SO₂
28 and PM_{2.5} emissions in Europe, induced by air quality policies resulting in implementation of abatement
29 measures since the 1970s, have caused a surface warming of $+0.45 \pm 0.11$ °C and increase of precipitation
30 $+13 \pm 0.8$ mm yr⁻¹ over Europe, compared to the scenario without such policies. While the temperature
31 increase is likely overestimated since the impact of increase in ammonium nitrate was not considered in this
32 study, the simulated European all-sky TOA radiative effect of the European air pollutant mitigation over the
33 period 1970-2009 is 2.5 times the change in global mean CO₂ radiative forcing over the same period (Myhre
34 et al., 2013b). Other studies found that the recent measures to reduce pollution over China have induced a
35 decrease of aerosols and increase of ozone over east China (Li et al., 2019a, 2020a) resulting in an overall
36 warming effect mainly due to the dominant effect of sulphate reductions in the period 2012-2017 (Dang and
37 Liao, 2019).

38 39 40 *6.6.3.2 Recently decided SLCF relevant global legislation*

41
42 International shipping emissions regulation: from January 2020, a new global standard, proposed by the
43 International Maritime Organisation, limits the sulphur content in marine fuels to 0.5% against previous
44 3.5% (IMO, 2016). This legislation is considered in the SSP5 and SSP2-4.5 and with a delay of few years in
45 SSP3-LowSLCF, SSP1-1.9, and SSP1-2.6, and in other SSP emission scenarios achieved by mid 21st
46 century. This global measure aims to reduce the formation of sulphate (and consequently PM_{2.5}) and largely
47 reduce the health exposure to PM_{2.5} especially over India, East China and coastal areas of Africa and the
48 Middle East (Sofiev et al., 2018). Sofiev et al. (2018) used a high spatial and temporal resolution chemistry-
49 climate model and estimated a net total ERF of $+71$ mW m⁻² associated with this measure and due to lower
50 direct aerosol cooling (-3.9 mW m⁻²) and lower cloud albedo (-67 mW m⁻²). This value, which corresponds to
51 80% decrease of the cooling effect of shipping induced by about 8 Tg of SO₂ of avoided emissions, is
52 consistent with older estimates which considered similar reduction of emitted sulphur. However, there is
53 considerable uncertainty in the indirect forcing since small changes in aerosols, acting as CCNs in clean
54 environment, can have disproportionately large effects on the radiative balance. Since sulphate is by far the
55 largest component of the radiative forcing (Fuglestad et al., 2008) and of surface temperature effect (Figure

1 6.16) due to ship emissions over short time scale, limiting the co-emitted SLCFs can not offset the warming
2 by sulphur reductions. The reduction of sulphur emission from shipping is assessed to lead to a slight
3 warming mainly due to aerosol-cloud interactions (*medium evidence, medium agreement*).

4
5 Kigali Amendment (UNEP, 2016): With the adoption of the Kigali Amendment to the Montreal Protocol
6 (UN, 1987) in 2016, parties agreed to the phasedown of HFCs, substances that are not ozone depleting but
7 are climate forcing agents (Papanastasiou et al., 2018). Baseline scenarios, in the absence of controls or only
8 pre-Kigali national legislation, projected increased use and emissions of HFCs. All recent baseline
9 projections are significantly higher than those used in the Representative Concentration Pathways (RCP)
10 scenarios ((Meinshausen et al., 2011) and Figure 6.18). There is *low confidence* that the high baseline
11 (assuming absence of controls, lack of technical progress, and high growth) as developed in Velders et al.
12 (2009), resulting in additional warming of about 0.5°C by 2100 (Xu et al., 2013; WMO, 2018), is plausible.
13 Evolution of HFC emissions along the baselines consistent with Velders et al. (2009) and Velders et al
14 (2015) would result in a global average warming, due to HFCs, relative to 2000, of about 0.1-0.12°C by
15 2050 and 0.35–0.5°C and 0.28–0.44°C by 2100, respectively (Xu et al., 2013). The baseline implementation
16 considered in SSP5-8.5 (see Section 6.7.1.1) is comparable to the lower bound of projections by Velders et
17 al. (2015) (Figure 6.18) and several other studies (Gschrey et al., 2011; Purohit and Höglund-Isaksson, 2017;
18 EPA, 2019; Purohit et al., 2020) and result in additional warming of 0.15-0.3°C by 2100 (Figure 6.22)
19 (*medium confidence*).

20
21 Efficient implementation of the Kigali Amendment and national and regional regulations has been projected
22 to reduce global average warming in 2050 by 0.05°–0.07°C (Klimont et al., 2017c; WMO, 2018) and by 0.2–
23 0.4°C in 2100 compared with the baseline (see Figure 2.20 of WMO, 2018). Analysis of SSP scenarios based
24 on an emulator (Section 6.7.3) shows a comparable mitigation potential of about 0.02–0.07°C in 2050 and
25 about 0.1–0.3°C in 2100 (Figure 6.22, SSP5-8.5 versus SSP1-2.6). Furthermore, the energy efficiency
26 improvements of cooling equipment alongside the transition to low global warming potential alternative
27 refrigerants for refrigeration and air-conditioning equipment could potentially increase the climate benefits
28 from the HFC phasedown under the Kigali Amendment (Shah et al., 2015; Höglund-Isaksson et al., 2017;
29 Purohit and Höglund-Isaksson, 2017; WMO, 2018). Purohit et al (2020) estimated that depending on the
30 expected rate of technological development, improving energy efficiency of stationary cooling technologies,
31 and compliance with Kigali Amendment could bring future global electricity-savings of more than 20% of
32 expected world's electricity consumption beyond 2050 or cumulative reduction of about 75-275 Gt CO₂eq
33 over the period 2018-2100 (*medium confidence*). This could potentially double the climate benefits of the
34 HFC phase-down of the Kigali Amendment as well as result in small air quality improvements due to
35 reduced air pollutant emissions from power sector, i.e., 8-16% reduction of PM_{2.5}, SO₂, NO_x (Purohit et al.,
36 2020).

37 38 39 6.6.3.3 Assessment of SLCF mitigation strategies and opportunities

40
41 There is a consensus in the literature that mitigation of SLCF emissions plays a central role in simultaneous
42 mitigation of climate change, air quality, and other development goals including SDG targets (UNEP and
43 WMO, 2011; Shindell et al., 2012b, 2017a, Rogelj et al., 2014b, 2018b; AMAP, 2015a; Haines et al., 2017a;
44 Klimont et al., 2017c; UNEP and CCAC, 2018; McCollum et al., 2018; Rafaj et al., 2018; UNEP, 2019).
45 There is less agreement in the literature with respect to the actual mitigation potential (or its potential rate of
46 implementation), necessary policies to trigger successful implementation, and resulting climate impacts.
47 Most studies agree that climate policies, especially those aiming to keep warming below 1.5-2°C, trigger
48 large SLCF mitigation co-benefits, (e.g., Rogelj et al., 2014b, 2018b), however, discussion of practical
49 implementation of respective policies and SDGs has only started (Haines et al., 2017a). Note that mitigation
50 scenarios outside of the SSP framework are assessed here while those within the SSPs are assessed in
51 Section 6.7.3.

52
53 Focusing on air quality, specifically addressing aerosol SLCFs, by introducing best available technology
54 reducing PM_{2.5}, SO₂, NO_x in most Asian countries within the 2030-2050 time frame (a strategy that has
55 indeed shown reduction in PM_{2.5} exposure in China) comes, in many regions, short of national regulatory

1 PM_{2.5} concentration standards (often set at 35 µg m⁻³ for annual mean) (UNEP, 2019). Similarly, global
2 studies (Rafaj et al., 2018; Amann et al., 2020) shows that strengthening current air quality policies,
3 addressing primarily aerosols and their precursors, will not allow to achieve WHO air quality guidelines
4 (annual average concentration of PM_{2.5} below 10 µg m⁻³) in many regions.
5

6 A multi-model study (four ESMs and six CTMs) found a consistent response to removal of SO₂ emissions
7 that resulted in a global mean surface temperature increase of 0.69°C (0.4-0.84°C), while mixed results for a
8 global BC-focused deep SLCF reduction, excluding SO₂ and CH₄ mitigation which remain like in the
9 baseline (see ECLIPSE in Figure 6.18), concluding about -0.022°C temperature reduction for the decade
10 2041-2050; this is derived from their estimate that mitigation of the non-CH₄ species contributed only about
11 10% of the global temperature reduction for the strategy where also CH₄ mitigation was included (-0.22°C
12 ±0.07°C) (Stohl et al., 2015). These results are consistent with studies analysing similar strategies using
13 emulators (Smith and Mizrahi, 2013; e.g., Rogelj et al., 2014b). Stohl et al. (2015) analysed also the impact
14 of BC-focused mitigation on air quality, estimating large scale regional reduction in PM_{2.5} mean
15 concentration from about 2% in Europe to 20% over India for the decade 2041-2050.
16

17 Local response to global reduction can be higher than the global temperature response in particular for
18 regions subjected to rapid changes. Hence, mitigation of rapid warming in the Arctic has been subject to
19 increasing number of studies (Sand et al., 2013a, 2016; Jiao et al., 2014; AMAP, 2015b, 2015a; Mahmood et
20 al., 2016; Christensen et al., 2019). Considering maximum technical mitigation potential for CH₄ globally
21 and an idealized strategy reducing key global anthropogenic sources of BC (about 80% reduction by 2030
22 and sustained thereafter) and precursors of O₃ was estimated to jointly bring a reduction of Arctic warming,
23 averaged over the 2041-2050 period, between 0.2 and 0.6°C (AMAP, 2015a; Sand et al., 2016). Stohl et al.
24 (2015) have estimated that a global SLCF mitigation strategy (excluding further reduction of SO₂) would
25 lead to about twice as high temperature reduction (-0.44°C (-0.39 to -0.49°C)) in the Arctic than the global
26 response to such mitigation.
27

28 While there is robust evidence that air-quality policies, resulting in reduction of aerosols and ozone, can be
29 beneficial for human health but can lead to ‘disbenefits’ for near-term climate change, such trade-offs in
30 response to climate mitigation policies is less certain (Shindell and Smith, 2019). Recent studies show that
31 very ambitious but plausible gradual phasing out of fossil-fuels in 1.5°C compatible pathways with little or
32 no overshoot, lead to a near-term future warming of less than 0.1°C, when considering associated emission
33 reduction of both warming and cooling species. This suggests that there may not be a strong conflict, at least
34 at the global scale, between climate and air-quality benefits in the case of a worldwide transition to clean
35 energy (Shindell and Smith, 2019; Smith et al., 2019). However, at the regional scale, the changes in
36 spatially variable emission and abundance changes might result in different responses including implications
37 for precipitation, monsoon, etc. (Chapter 8), especially over South Asia (e.g., Wilcox et al., 2020).
38

39 Decarbonization of energy supply and end use sectors is among key pillars of any ambitious climate
40 mitigation strategy and it would result in improved air quality owing to associated reduction of co-emitted
41 SLCF emissions (McCollum et al., 2013; Rogelj et al., 2014b; e.g., Braspenning Radu et al., 2016; Rao et al.,
42 2016; Stechow et al., 2016; Lelieveld et al., 2019; Shindell and Smith, 2019). Regional studies (Lee et al.,
43 2016; Shindell et al., 2016; Chen et al., 2018; Li et al., 2018b), where significant CO₂ reductions were
44 assumed for 2030 and 2050, show consistently reduction of PM_{2.5} and ozone concentrations resulting in
45 important health benefits. However, these improvements are not sufficient to bring PM_{2.5} levels in agreement
46 with the WHO air quality guideline in several regions. Amann et al. (2020) and UNEP (2019) highlight that
47 only combination of strong air quality, development, and climate policies, including societal transformations,
48 could pave the way towards achievement of such target at a regional and global level.
49

50 At a global level, Rao et al. (2016) showed that climate policies, compatible with Copenhagen pledges and a
51 long-term CO₂ target of 450 ppm, result in important air quality benefits, reducing the share of global
52 population exposed to PM_{2.5} levels above the WHO Tier 1 standard (35 µg m⁻³) in 2030 from 21% to 5%.
53 The impacts are similar to a strong air quality policy but still leaving large parts of population, especially in
54 Asia and Africa, exposed to levels well above WHO air quality guideline level of 10 µg m⁻³. The latter can
55 be partly alleviated by combining such climate policy with strong air quality policy. Shindell et al. (2018)

1 analysed more ambitious climate mitigation scenarios than Rao et al. (2016) and highlighted the
2 opportunities to improve air quality and avert societal effect associated with warmer climate by accelerated
3 decarbonization strategies. Most climate mitigation strategies compatible with limiting global warming to
4 well below 2°C, rely on future negative CO₂ emissions postponing immediate reduction. Alternatively, a
5 faster decarbonization could allow to achieve a 2°C goal without negative CO₂ emissions and with currently
6 known and effectively applied emission control technologies would have also immediate and significant air
7 quality benefits, reducing premature deaths worldwide (Shindell et al., 2018). For a 2°C compatible pathway,
8 Vandyck et al. (2018) estimated 5% and 15% reduction in premature mortality due to PM_{2.5} in 2030 and
9 2050, respectively, compared to reference scenarios.

10
11 There is robust evidence that reducing atmospheric CH₄ will benefit climate and improve air quality through
12 near-surface O₃ reduction (Fiore et al., 2015; Shindell et al., 2017b) and wide agreement that strategies
13 reducing CH₄ offer larger (and less uncertain) climate benefits than policies addressing BC (Smith and
14 Mizrahi, 2013; Rogelj et al., 2014b, 2018b; Stohl et al., 2015; e.g., Christensen et al., 2019; Shindell and
15 Smith, 2019). SR1.5 (Rogelj et al., 2018b) highlighted the importance of CH₄ mitigation in limiting warming
16 to 1.5 °C in addition to net zero CO₂ emission by 2050. Implementation of the identified maximum technical
17 mitigation (MTFR) potential for CH₄ globally, estimated at nearly 50% reduction (or 205 Tg CH₄ in 2050) of
18 anthropogenic emissions from the baseline, would lead to a reduction in warming, calculated as the
19 differences between the baseline and MTFR scenario, for the 2036-2050 period of about 0.20±0.02°C
20 globally (AMAP, 2015b). Plausible levels of methane mitigation, achieved with proven technologies, can
21 increase the feasibility of achieving the Paris Agreement goal through slightly slowing down the pace of CO₂
22 reductions (but not changing the final CO₂ reduction goal) while this benefit is enhanced by the indirect
23 effects of methane mitigation on ozone levels (Collins et al., 2018). Addressing methane mitigation appears
24 even more important in view of recently observed growth in atmospheric concentrations that is linked to
25 increasing anthropogenic emissions (see Section 5.1.1.).

26
27 Neither ambitious climate change policy nor air quality abatement policy can automatically yield co-benefits
28 without integrated policies aimed at co-beneficial solutions (Zusman et al., 2013; Schmale et al., 2014b;
29 Melamed et al., 2016), particularly in the energy generation and transport sectors (Rao et al., 2013;
30 Thompson et al., 2016; Shindell et al., 2018; Vandyck et al., 2018). Integrated policies are necessary to yield
31 multiple benefits of mitigating climate change, improving air quality, protecting human health, and achieving
32 several Sustainable Development Goals.

33
34
35 **[START BOX 6.2 HERE]**

36 **BOX 6.2: SLCF Mitigation and Sustainable Development Goals (SDG) opportunities**

37 Striving to achieve air quality and climate targets will bring significant SLCFs reductions. These reductions
38 contribute first and foremost to attainment of SDGs targeting improved human health and sustainable cities
39 (SDG 3 and 11), specifically related to PM exposure (goals 3.9 and 11.6) (Lelieveld, 2017; Amann et al.,
40 2020) but also access to affordable and clean energy, responsible consumption and production, climate, as
41 well as reducing nutrient losses and consequently protect biodiversity (SDG 7, 12, 13, 14, and 15) (UNEP,
42 2019; Amann et al., 2020). Furthermore, declining SLCF emissions will result in reduced crop losses (SDG
43 2; zero hunger) due to decrease of ozone exposure (Feng and Kobayashi, 2009; Ainsworth et al., 2012;
44 Emberson et al., 2018).

45
46 However, the design of suitable policies addressing these SDGs can be difficult because of the complexity
47 linking emissions to impacts on human health, ecosystem, equity, infrastructure, and costs. Beyond the fact
48 that several species are co-emitted, interlinkage between species, for example through atmospheric
49 chemistry, can weaken the benefit of emissions reduction efforts. An illustration lies in the recent (2013-
50 2017) reduction of aerosols over China (Silver et al., 2018; Zheng et al., 2018b) resulting from the strategy to
51 improve air quality (“Clean Air Action”), has successfully reduced the level of PM_{2.5} but has led to a
52 concurrent increase in surface ozone, partly due to declining heterogeneous interactions of O₃ precursors
53 with aerosols (Li et al., 2019a; Yu et al., 2019). This side effect on ozone has been addressed since then by
54
55

1 amending the legislation to target NMVOC sources, especially solvent use. Complex interactions between
2 anthropogenic and biogenic vapours are also at play and reduction of certain SLCFs could possibly promote
3 new particle formation from organic vapours (e.g., Lehtipalo et al., 2018). Finally, a recent example of this
4 complexity is the mixed effects on ozone pollution induced by NO_x decrease during the COVID-19
5 pandemic (see Cross-Chapter Box 6.1). Thus, the climate and air pollution effects of policies depend
6 strongly on choice of regulated compounds and the degree of reduction. Such policies have to be informed
7 by strong science support, including for example multi-model analyses, e.g., HTAP (UNECE, 2010), AMAP
8 (AMAP, 2015a, 2015b), based on global and regional CCMs, essential to capture the complexity and inform
9 the policy development process.

10 In addition, pursuing SDG objectives, apparently decoupled from air pollution, such as improved waste
11 management, access to clean energy, or improved agricultural practices, would also stimulate and lead to
12 mitigation of SLCFs (Box 6.2, Figure 1). Amann et al. (2020) shows that a global strategy to achieve the
13 WHO air quality guidelines, cannot only rely on air pollution control but also on a combination of SDG
14 aligned policies. Such actions would include energy efficiency improvements, increased use of renewables,
15 reduction of CH₄ from waste management and agriculture, and CO₂ and CH₄ due to lower fossil fuel
16 consumption, resulting in climate co-benefits. Consideration of SDGs including local air quality co-benefits,
17 creates an opportunity to support and gain acceptance for ambitious climate mitigation (Jakob and Steckel,
18 2016; Stechow et al., 2016; Vandyck et al., 2018). Such near-term policies targeting SDGs and air quality
19 would enable longer term transformations necessary to achieve climate goals (Chapter 17, WGIII).

20
21 **[END BOX 6.2 HERE]**
22
23

24 In summary, there is *high confidence* that effective decarbonization strategies could lead to air quality
25 improvements but are not sufficient to achieve, in the near term, air quality WHO guideline values set for
26 fine particulate matter), especially in parts of Asia and in some highly polluted regions. Additional policies
27 (e.g., access to clean energy, waste management) envisaged to attain Sustainable Development Goals (SDG)
28 bring complementary SLCF reduction (*high confidence*). Sustained methane mitigation, wherever it occurs,
29 stands out as an option that combines near and long-term gains on surface temperature (*high confidence*) and
30 leads to air pollution benefit by reducing globally the surface ozone level (*high confidence*).
31

32
33 **[START CROSS-CHAPTER BOX 6.1 HERE]**
34

35 **Cross-Chapter Box 6.1: Implications of COVID-19 restrictions for emissions, air quality and** 36 **climate**

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44
45
46

47 In response to the outbreak of COVID-19 (officially the severe acute respiratory syndrome–coronavirus 2 or
48 SARS-CoV-2), which was declared a pandemic on March 11 2020 by the World Health Organization
49 (WHO), regulations were imposed by many countries to contain the spread of COVID-19. Restrictions were
50 implemented on the movement of people, such as closing borders or requiring the majority of population to
51 stay at home, for periods of several months. This Cross Chapter Box assess the influence of the COVID-19
52 containment on short-lived climate forcers (SLCFs) and long-lived greenhouse gases (LLGHGs) and related
53 implications for the climate. Note that this assessment was developed late in the AR6 WGI process and is
54 based on the available emerging literature.
55

Emissions

Global fossil CO₂ emissions are estimated to have declined by 7% (*medium confidence*) in 2020 compared to 2019 emissions, with estimates ranging from 5.8% to 13.0% based on various combinations of data on energy production and consumption, economic activity and proxy activity data for emissions and economic activity and proxy activity data for emissions and their drivers

(Forster et al., 2020; Friedlingstein et al., 2020; Le Quéré et al., 2020; Liu et al., 2020).

However, the concentration of atmospheric CO₂ continued to grow in 2020 compared to previous years (Dlugokencky and Tans, 2021). Given the large natural inter-annual variability of CO₂ (Section 5.2.1), and the small expected impact of emissions in the CO₂ growth rate, there were no observed changes in CO₂ concentration that could be attributed to COVID-19 containment (Chevallier et al., 2020; Tohjima et al., 2020).

Global daily CO₂ emissions from fossil fuel sources had a maximum decline of 17% in early April, compared with the mean 2019 levels, and coinciding with the global peak pandemic lockdown (Le Quéré et al., 2020). The reductions in CO₂ emissions in 2020 were dominated by the drop in emissions from surface transport followed, in order of absolute emission reductions, by industry, power, and aviation (Le Quéré et al., 2020; Liu et al., 2020). Residential emissions showed little change (Liu et al., 2020) or rose slightly (Forster et al., 2020; Le Quéré et al., 2020). Aviation had the biggest relative drop in activity accounting. CO₂ emissions due to land-use (based on early and uncertain evidence on deforestation and forest fires) were higher than average in 2020 (Amador-Jiménez et al., 2020).

Using similar methodologies, (Forster et al., 2020) assembled activity data and emissions estimates for other greenhouse gases and aerosols and their precursors. Anthropogenic NO_x emissions, which are largely from the transport sector, are estimated to have decreased by a maximum of 35% in April (*medium confidence*). Species whose emissions are dominated by other sectors, such as CH₄ and NH₃ from agriculture, saw smaller reductions.

Abundances and air quality

Owing to the short atmospheric lifetimes of SLCFs relevant to air quality, changes in their concentrations were detected within few days after lockdowns had been implemented (e.g., Bauwens et al., 2020b; Gkatzelis et al., 2021; Shi et al., 2021; Venter et al., 2020b). The COVID-19 driven economic slowdown has illustrated how complex the relationship is between emissions and air pollutant concentrations due to non-linearity in the atmospheric chemistry leading to secondary compound formation (see also Section 6.1 and Box 6.1) (Kroll et al., 2020).

Several studies have examined the effect of COVID-19 containment on air quality, showing that multi-year datasets with proper statistical/modelling analysis are required to discriminate the effect of meteorology from that of emission reduction (Dhaka et al., 2020; Li et al., 2020b; Wang et al., 2020; Zhao et al., 2020b). Accounting for meteorological influences and with increasing stringency index, the median observed change in NO₂ decreased from -13% to -48%, and in PM_{2.5} decreased from -10% to -33%, whereas the median change in O₃ increased from 0% to 4% (Gkatzelis et al., 2021). The latter can be explained by the decrease of NO emissions that titrate ozone in specific highly polluted areas, leading to the observed increase in surface O₃ concentration in cities (Huang et al., 2020a; Le et al., 2020; Sicard et al., 2020).

The temporary decrease of PM_{2.5} concentrations should be put in perspective of the sustained reduction (estimated at 30% to 70%), which could be achieved by implementing policies addressing air quality and climate change (see Section 6.6.3). Such sustained reductions can lead to multiple benefits and simultaneously achieve several SDGs (Section 6.6.3). These policies would also result in reduction of ground-level ozone by up to 20% (see Section 6.7.1.3).

Except for ozone, temporary improvement of air quality during lockdown periods was observed in most regions of the world (*high confidence*), resulting from a combination of inter-annual meteorological variability and impact of COVID-19 containment measures (*high confidence*). Estimated air pollution

1 reductions associated with lockdown periods are lower than what can be expected from integrated mitigation
2 policy leading to lasting reductions (*medium confidence*).

3 4 **Radiative forcings**

5
6 COVID-19 related emission changes primarily exerted effective radiative forcing (ERF) through reduced
7 emission rates of CO₂ and methane, altered abundance of short-lived climate forcers (SLCFs), notably O₃,
8 NO₂ and aerosols, and through other changes in anthropogenic activities, notably a reduction in the
9 formation of aviation-induced cirrus clouds.

10
11 Forster et al. (2020) combined the FaIR emulator (see Cross-Chapter Box 7.1) with emission changes for a
12 range of species, relative to a continuation of Nationally Determined Contributions (Rogelj et al., 2017).
13 They found a negative ERF from avoided CO₂ emissions that strengthens through 2020, to -0.01 Wm⁻².
14 During the spring lockdown, they found a peak positive ERF of 0.1 Wm⁻² from loss of aerosol-induced
15 cooling, and a peak negative ERF of -0.04 Wm⁻² from reductions in tropospheric ozone (from reduced
16 photochemical production via NO_x). Overall, they estimated a net ERF of +0.05 Wm⁻² for spring 2020,
17 declining to +0.025 Wm⁻² by the end of the year.

18
19 Gettelman et al. (2021) extended Forster et al. (2020) results using two ESMs, and found a spring peak
20 aerosol-induced ERF ranging from 0.12 to 0.3 Wm⁻², depending on the aerosol parameterization. They also
21 estimated an ERF of -0.04 Wm⁻² from loss of contrail warming. Overall, they report a peak ERF of 0.04 to
22 0.2 Wm⁻², and a subsequent decline to around half the peak value.

23
24 Two independent ESM studies, Weber et al. (2020) and Yang et al. (2020), found consistent results in time
25 evolution and component contributions, but included fewer forcing components.

26
27 The available studies are in broad agreement on the sign and magnitude of contributions to ERF from
28 COVID-19 related emission changes during 2020. The peak global mean ERF, in spring 2020,
29 was [0.025-0.2] Wm⁻² (*medium confidence*), composed by a positive forcing from aerosol-climate
30 interactions of [0.1-0.3] Wm⁻², and negative forcings from CO₂ (-0.01 Wm⁻²), NO_x (-0.04 Wm⁻²) and contrail
31 cirrus (-0.04 Wm⁻²) (*low evidence, medium agreement*). By the end of 2020, the ERF was at half the peak
32 value (*medium confidence*).

33 34 **Climate responses**

35
36 Changes in atmospheric composition due to COVID-19 emissions reductions are not thought to have caused
37 a detectable change in global temperature or rainfall in 2020 (*high confidence*). A large ensemble of Earth
38 System Model (ESM) simulations show an ensemble average reduction in Aerosol Optical Depth (AOD) in
39 some regions, notably eastern and southern Asia (Fyfe et al., 2021). This result is supported by observational
40 studies finding decreases in optical depth in 2020 (Gkatzelis et al., 2021; Ming et al., 2021; van Heerwaarden
41 et al., 2021), which may have contributed to observed increases in solar irradiance (van Heerwaarden et al.,
42 2021) or solar clear-sky reflection (Ming et al., 2021).

43
44 Model simulations of the response to COVID-19 emissions reductions indicate a small warming of global
45 surface air temperature (GSAT) due to a decrease in sulphate aerosols (Forster et al., 2020; Fyfe et al., 2021),
46 balanced by cooling due to an ozone decrease (Forster et al., 2020; Weber et al., 2020), black carbon
47 decrease (Weber et al., 2020) and CO₂ decrease. It is noted that observational studies report little SO₂
48 change, at least locally near the surface (Shi et al., 2021) and do not correlate with emission inventory based
49 changes (Gkatzelis et al., 2021). One study suggests a small net warming while another using idealized
50 simulations suggests a small cooling (Weber et al., 2020). Simulated GSAT and rainfall changes are
51 unlikely to be detectable in observations (Samset et al., 2020; Fyfe et al., 2021) (*high confidence*).

52
53 Multi-model ESM simulations based on a realistic COVID-19 containment forcing scenario (Forster et al.,
54 2020) indicate a model mean reduction in regional AOD but no discernible response in GSAT (Cross-
55 Chapter Box 6.1 Figure 1).

Scenarios of future emissions following COVID-19 disruption have been examined under a range of hypothetical assumptions (Forster et al., 2020).

[START CROSS-CHAPTER BOX 6.1, FIGURE 1 HERE]

Cross-Chapter Box 6.1, Figure 1: Emission reductions and their effect on aerosols and climate in response to COVID-19. Estimated reductions in emissions of CO₂, SO₂ and NO_x are shown in panel (a) based on reconstructions using activity data (updated from Forster et al., 2020). Eight Earth System Models (ESMs) performed multiple ensemble simulations of the response to COVID-19 emissions reductions forced with these assumed emission reductions up till August 2020 followed by a constant continuation near the August value to the end of 2020. Emission reductions were applied relative to the SSP2-4.5 scenario. Panel (b) shows ESM simulated AOD at 550nm (only seven models reported this variable). Panel (c) shows ESM simulated GSAT anomalies during 2020; curves denote the ensemble mean result for each model with shading used for ±1 standard deviation for each model. ESM data from these simulations (“ssp245-covid”) is archived on the Earth System Grid CMIP6 database. Uncertainty is represented using the simple approach: No overlay indicates regions with high model agreement, where ≥80% of models agree on sign of change; diagonal lines indicate regions with low model agreement, where <80% of models agree on sign of change. For more information on the simple approach, please refer to the Cross-Chapter Box Atlas.1.

[END CROSS-CHAPTER BOX 6.1, FIGURE 1 HERE]

[END CROSS-CHAPTER BOX 6.1 HERE]

6.7 Future projections of Atmospheric Composition and Climate response in SSP scenarios

This section assesses the 21st century projections of SLCF emissions, abundances, and responses in terms of climate and air quality following the SSPs (Riahi et al., 2017; Gidden et al., 2019) (also see Chapter 1, Cross-Chapter Box 1.5, and Section 1.6.1.3). The future evolution of atmospheric abundances and the resulting climate and AQ responses is driven mainly by anthropogenic emissions and by natural emissions modulated by chemical, physical and biological processes as discussed in Sections 6.2 and 6.3. Like the RCP scenarios used in AR5, the SSP emission scenarios consider only direct anthropogenic (including biomass burning) emissions and do not project natural emission changes due to climate or land-use changes; ESMs intrinsically consider these biogeochemical feedbacks to varying degrees (Section 6.4.5). We rely on future projections based on CMIP6 ESMs with comprehensive representation of chemistry, aerosol microphysics, and biospheric processes that participated in the ScenarioMIP (O’Neill et al., 2016) and AerChemMIP (Collins et al., 2017). However, due to high computational costs of running coupled ESMs, they cannot be used for quantifying the contributions from individual species, regions and sectors and across the scenarios. Therefore, reduced complexity models (see Box 1.3 and Cross-Chapter Box 7.1), which represent chemistry and complex ESM interactions in parameterized forms updated since the AR5, are also applied here.

6.7.1 Projections of Emissions and Atmospheric Abundances

6.7.1.1 SLCF Emissions and atmospheric abundances

The trajectory of future SLCF emissions is driven by the evolution of socioeconomic drivers described in 1.6.1.1 but dedicated, SSP-specific, air pollution policy storylines can change the regional and global trends (Rao et al., 2017). Additionally, assumptions about urbanization (Jiang and O’Neill, 2017) will affect spatial distribution of emissions and consequently air quality.

1 Growing urbanization worldwide has strongly modified the spatial distribution and intensity of SLCF
2 emissions. The effect and extent of urbanization on air pollution and other emissions species are captured
3 within Integrated Assessment Models (IAM) at varying levels of complexity. In most cases, models use a
4 combination of proxies and assumptions of end-use efficiency and technological improvement assumptions
5 to estimate emissions species arising from rural-to-urban migration and population growth within cities,
6 utilizing quantifications of urbanization for the SSPs (Jiang and O'Neill, 2017). In addition, spatial patterns
7 of future rural and urban population growth, migration, and decline have been quantified for the SSPs using
8 a gravity model (Jiang and O'Neill, 2017). However, linking these spatial patterns with IAM regional
9 emissions pathways is still an ongoing area of study and has not yet been represented in spatial emissions
10 estimates provided by IAMs (Riahi et al., 2017; Gidden et al., 2019; Feng et al., 2020). As described in Feng
11 et al. (2020), spatial emissions estimates derived for CMIP6 are largely a product of existing spatial patterns
12 of population, but do not vary dynamically in future patterns. To the extent urbanization is accounted for in
13 gridded emissions, IAM native region resolution (varying, for example, from 11 world regions to more than
14 30, depending on the model) provides urbanization-based dynamics. Despite the interest of studying the
15 effect of well planned, densely populated urban centres, which can help to maximize the benefits of
16 agglomeration, by providing proximity to infrastructure and services, opportunity for energy saving, and
17 provide a frame for air quality control, IAM realizations of SSPs are not sufficient to assess this effect. The
18 opportunities and risks associated with this rapid urbanization for SLCF emissions and air quality are
19 analysed in the Chapter 6 of the WGII report and Chapter 8 of the WGIII report.

21 All the RCP trajectories started in 2005 relying on the assumption that economic growth will bring rapid
22 strengthening of air pollution legislation effectively reducing emissions of non-methane SLCFs (e.g.,
23 Chuwah et al., 2013). While in the long term such trends are expected if more ambitious air pollution control
24 goes on par with the economic growth. The near term developments, however, might be much more diverse
25 across the regions and species as has been observed in the last three decades (Amann et al., 2013; Rafaj et
26 al., 2014; Rafaj and Amann, 2018; Ru et al., 2018), especially in several fast-growing economies, leading to
27 the difference between CMIP6 historical estimates for the post 2000 period (Hoesly et al., 2018) and those
28 used in RCPs (Figure 6.18). Since several SLCFs are also air pollutants, the narrow range of the RCP
29 emissions trajectories in the future allowed for only limited analysis of near-future air quality (Amann et al.,
30 2013; Chuwah et al., 2013; Von Schneidmesser et al., 2015). However, the range of storylines in the SSPs
31 lead to a wider range of assumed pollution control policies in the SSPs (Rao et al., 2017; Riahi et al., 2017).
32 In SSP1 and SSP5, strong air quality policies are assumed to minimize adverse effects of pollution on
33 population and ecosystems. In SSP2, a medium pollution control, with lower than current policy targets, is
34 considered. Only weak, regionally varied, air pollution policies are applied in the SSP3 and SSP4. Additional
35 climate policies introduced to reach defined radiative forcing targets will also affect SLCF emissions. The
36 SSP SLCF emission trajectories (Rao et al., 2017; Gidden et al., 2019) assume a long term coupling of
37 economic growth and specific emission indicators, such as, sectoral emission densities. The pace of change
38 varies across regions and SSPs resulting in a wider range of future air pollutants evolution (Figure 6.18)
39 reflecting the differences in assumed level of air pollution controls across the regions (Figure 6.19). At the
40 end of the century, the SSP scenarios range is about four times that of RCP for SO₂ and NO_x, two to four
41 times for BC and NMVOC, and up to three times for CO and OC, while indicating slightly smaller range
42 than RCPs for CH₄ (Figure 6.20). The originally developed SSP scenarios (Rao et al., 2017) have been
43 harmonized with the CMIP6 historical emissions (Hoesly et al., 2018) and include updated SO₂ emissions to
44 account for the recent decline in China (Gidden et al., 2019).

46 All SSP scenarios (Figure 6.18), except SSP3-7.0, project decline in global total emissions for all SLCFs by
47 the end of the 21st century, except for ammonia and for HFCs where Kigali Amendment is not included (see
48 Section 6.6.3.2). Similar to RCPs, ammonia emissions continue to increase in most SSPs, except SSP1 and
49 SSP2, accounting for the expected growth in food demand and a general lack of effective policies targeting
50 agricultural emissions. Additionally, mitigation potential for NH₃ is generally smaller than for other species
51 owing to fugitive and widely distributed sources (Pinder et al., 2006; Klimont and Winiwarter, 2015;
52 Mohankumar Sajeev et al., 2018; Sajeev et al., 2018). Most significant changes of SLCF emissions in the
53 near and long term compared to present day are expected for SO₂ owing to ever more stringent (and
54 enforced) legislation in China's power sector, extended recently to industrial sources (Zheng et al., 2018b;
55 Tong et al., 2020), declining coal use in most SSPs, recently announced stricter emission limits for power

1 sector in India, and reduction of sulphur content of oil fuel used in international shipping from 2020 (IMO,
2 2016). For the lower forcing targets (e.g., SSP1-2.6), the SO₂ trajectories are similar to the RCPs resulting in
3 over 50% to 90% decline by 2050 and 2100, respectively, while for the scenarios with no climate policies,
4 the SSPs show large spread even at the end of the century.

5
6
7 **[START FIGURE 6.18 HERE]**

8
9 **Figure 6.18: Global anthropogenic and biomass burning short-lived climate forcer (SLCF) and carbon dioxide**
10 **(CO₂) emissions from 1850 to 2100 and HFC emissions from 1990 to 2100.** Emissions for the Coupled
11 Model Intercomparison Project Phase 6 (CMIP6) for the period 1850-2014 are based on Hoesly et al.
12 (2018) and van Marle et al. (2017), emissions for CMIP5 for the period 1850-2005 are from Lamarque et
13 al. (2010), CO₂ emissions are from EDGAR database (Crippa et al., 2020); CH₄ and HFCs are from
14 (Crippa et al., 2019); and air pollutants are from (EC-JRC / PBL, 2020), Höglund-Isaksson (2012) and
15 Klimont et al. (2017b) for ECLIPSE. Projections originate from the Shared Socio-Economic Pathway
16 (SSP) database (Riahi et al., 2017; Rogelj et al., 2018a; Gidden et al., 2019), Representative
17 Concentration Pathway (RCP) database (van Vuuren et al., 2011), GAINS (CLE – current legislation
18 baseline, KA – Kigali Amendment, MTR – maximum technical mitigation potential) for HFCs (Pyrohit
19 et al., 2020), Velders et al. (2015), ECLIPSE (Stohl et al., 2015). Further details on data sources and
20 processing are available in the chapter data table (Table 6.SM.1).

21
22 **[END FIGURE 6.18 HERE]**

23
24
25 **[START FIGURE 6.19 HERE]**

26
27 **Figure 6.19: Regional anthropogenic and biomass burning short-lived climate forcer (SLCF) emissions from**
28 **1850 to 2100.** Emissions for the Coupled Model Intercomparison Project Phase 6 (CMIP6) for the period
29 1850-2014 are based on Hoesly et al. (2018) and van Marle et al. (2017) and emissions for CMIP5 for
30 the period 1850-2005 are from Lamarque et al. (2010). Projections originate from the Shared Socio-
31 Economic Pathway (SSP) database (Riahi et al., 2017; Rogelj et al., 2018a; Gidden et al., 2019) and
32 Representative Concentration Pathway (RCP) database (van Vuuren et al., 2011). Further details on data
33 sources and processing are available in the chapter data table (Table 6.SM.1).

34
35 **[END FIGURE 6.19 HERE]**

36
37
38 Until mid-century, SSP3-7.0 and SSP5-8.5 scenarios project no reduction in NO_x emissions at the global
39 level with decline in most OECD countries and East Asia, driven by existing legislation in power, industry,
40 and transportation, e.g., Tong et al. (2020), and continued increase in the rest of the world (Figure 6.18 and
41 6.19). Towards the end of the century, similar trends continue in SSP3-7.0 while emissions in SSP5 decline
42 strongly owing to faster technological progress and stronger air quality action (Rao et al., 2017; Riahi et al.,
43 2017). By 2100, the ‘Regional Rivalry’ (SSP3) scenario emissions of NO_x (and most other SLCFs, except
44 ammonia) are typically twice as high as next highest SSP projection, both at the global (Figure 6.18) and
45 regional level (Figure 6.19). In emission pathways consistent with Paris Agreement goals, NO_x drops,
46 compared to 2015, by 50% in SSP1-2.6 and by 65% in SSP1-1.9 by 2050, and by 2100 is reduced by about
47 70%, resulting in global emission levels comparable to 1950s and below the RCP range. Similar reductions
48 are projected in climate mitigation pathways at the regional level, except Africa (less than 50% decline) due
49 to high share of biomass emissions as well as strong growth in population and fossil fuel use. The trends in
50 anthropogenic and biomass burning emissions for other ozone precursors (NMVOC, CO) are similar to that
51 of NO_x.

52
53 An additional scenario, based on the SSP3-7.0, has been designed specifically to assess the effect of a strong
54 SLCF emission abatement and is called SSP3-7.0-lowNTCF in the literature (Collins et al., 2017; Gidden et
55 al., 2019). It has been applied in the modelling studies (e.g., AerChemMIP) with or without consideration of
56 additional CH₄ reduction and we refer here to these scenarios, respectively, as SSP3-7.0-lowSLCF-lowCH₄
57 or SSP3-7.0-lowSLCF-highCH₄. In these scenarios, aerosols, their precursors, and non-methane tropospheric

1 ozone precursors are mitigated by applying the same emission factors as in SSP1-1.9.

2
3 For global methane emissions, the range is similar for SSPs and RCPs over the entire century (Figure 6.18),
4 with highest projections in SSP3-7.0 (slightly below RCP8.5) estimating doubling of the current emissions
5 and a reduction of about 75% by 2100 in scenarios consistent with 1.5-2°C targets; similar as in RCP2.6. At
6 the regional level, the evolution of CH₄ emissions in climate mitigation scenarios is comparable to RCPs but
7 there are significant differences for some regions with respect to high-emission scenarios. Especially,
8 projection for East Asia differ significantly, the highest SSP3-7.0 is about half of the highest RCP by 2100
9 (Figure 6.19) which is due to much lower projections of coal use in China driven largely by the last decade
10 efforts to combat poor air quality. At the same time, the SSP scenarios without climate mitigation project
11 faster growth in CH₄ emissions in Africa, Middle East, and Latin America (Figure 6.19) driven by
12 developments in agriculture, oil and gas sectors, and, especially in Africa, waste management. There are
13 significant differences in the assessment and feasibility of rapid CH₄ mitigation. Höglund-Isaksson et al.
14 (2020) review most recent studies and assess feasibility of rapid widespread mitigation, concluding that
15 significant (over 50%) reductions are attainable but feasibility of such reductions could be constrained in the
16 short term due to locked capital. This might have implications for near term evolution assumed in, for
17 example, SSP1-1.9 or SSP3-lowSLCF-lowCH₄, where emissions drop very quickly due to fast
18 decarbonization and reductions in agriculture. Such high reduction potential in agriculture has been also
19 assumed in other studies (Lucas et al., 2007; Harmsen et al., 2019) but is questioned by Höglund-Isaksson et
20 al. (2020) who indicate that widespread implementation (within decades) of policies bringing about
21 institutional and behavioural changes would be important for transition towards very low CH₄ emissions
22 from livestock production.

23
24 Global emissions of carbonaceous aerosols are projected to decline in all SSP scenarios (Figure 6.18) except
25 SSP3-7.0. In that scenario, which also has much higher emissions than any of the RCPs, about half of the
26 anthropogenic BC originates from cooking and heating on solid fuels, mostly in Asia and Africa (Figure
27 6.19), where only limited progress in access to clean energy is achieved. Slow progress in improving waste
28 management, high coal use in energy and industry, and no further progress in controlling diesel engines in
29 Asia, Africa, and Latin America contributes most of the remaining emissions resulting in about 90% of
30 anthropogenic BC emitted in non-OECD world by 2100 in SSP3-7.0. Similar picture emerges for OC but
31 with larger importance of waste management sector and biomass burning and lower impact of transportation
32 and industry developments. Since scenarios compliant with Paris Agreement goals (SSP1-1.9 or SSP1-2.6;
33 see Section 1.6.1) include a widespread access to clean energy already by 2050, the global and regional
34 emissions of BC decline by 70-75% by 2050 and 80% by 2100. The decline in residential sector (about 90%
35 by 2050 and over 95% by 2100) is accompanied by a strong reduction in transport (over 98%) and
36 decarbonization of industry and energy sector. About 50% of remaining BC emissions in SSP1-1.9 or SSP1-
37 2.6 is projected to originate from waste and open biomass burning of which open burning of residues
38 represent a significant part. Some studies suggest this might be pessimistic as, for example, efficient waste
39 management (consistent with SDG goals) could potentially eliminate open burning of solid waste (Gómez-
40 Sanabria et al., 2018), which accounts for over 30% of BC emissions in SSP1-1.9 in 2050 or 2100.

41
42 The SSP scenarios draw on the HFCs projections developed by Velders et al. (2015) considering, in climate
43 mitigation scenarios, the provisions of the Kigali Amendment (2016) to the Montreal Protocol leading to
44 phasedown of HFCs (Papanastasiou et al., 2018) (see Section 6.6.3.2). The SSP scenarios without climate
45 mitigation (e.g., SSP3-7.0, SSP5-8.5) show a range of 3.2-5.3 Gt CO₂-eq yr⁻¹ in 2050 and about 4-7.2
46 Gt CO₂-eq yr⁻¹ by 2100 while in deep climate mitigation scenarios (e.g., SSP1-1.9), consistent with the 1.5-
47 2°C targets, they are expected to drop to 0.1-0.3 and 0.1-0.35 GtCO₂-eq yr⁻¹, respectively (Figure 6.19). In
48 SSP1-1.9, the extent of reduction and its pace is more ambitious than current estimates of the effect of fully
49 implemented and enforced Kigali Amendment (Figure 6.19) (Höglund-Isaksson et al., 2017; Purohit and
50 Höglund-Isaksson, 2017). The best representation of the HFC emission trajectories in the SSP framework
51 compliant with the Kigali Amendment is the SSP1-2.6 and the baseline (including only pre-Kigali national
52 legislation; 6.6.3) is best represented by SSP5-8.5 (see Figure 6.19). However, since HFC emissions in SSPs
53 were developed shortly after the Kigali Amendment had been agreed, none of these projections represents
54 accurately the HFC emission trajectory corresponding to the phase-out emission levels agreed to in the
55 Kigali Amendment (Meinshausen et al., 2020), leading to *medium confidence* in the assessment of the

1 benefits of the Kigali Amendment when using SSP projections for HFCs.

2
3 The SSP SLCF trajectories reflect the effect of recent legislation and assumed evolution thereof in the longer
4 term, however, they do not necessarily reflect the full mitigation potential for several SLCFs, within
5 particular SSPs (Figure 6.18), that could be achieved with air quality or SDG targeted policies (Amann et al.,
6 2013; Rogelj et al., 2014a; Haines et al., 2017b; Klimont et al., 2017c; Rafaj and Amann, 2018; Shindell
7 et al., 2018; Tong et al., 2020). Such policies could bring more rapid mitigation of SLCFs, independent of the
8 climate strategy (also see Section 6.6.3).

9
10 The projections of future SLCF abundances typically follow their emissions trajectories except for SLCFs
11 that are formed from precursor reactions (e.g., tropospheric ozone) or are influenced by biogeochemical
12 feedbacks (See 6.2.2 and 6.4.5). According to multi-model CMIP6 simulations, total column ozone
13 (reflecting mostly stratospheric ozone) is projected to return to 1960s values by the middle of the 21st
14 century under the SSP2-4.5, SSP3-7.0, SSP4-3.4, SSP4-6.0 and SSP5-8.5 scenarios (Keeble et al., 2021).
15 ESMs project increasing tropospheric ozone burden over the 2015-2100 period for the SSP3-7.0 scenario
16 (Figure 6.4) (Griffiths et al., 2020), there is, however, a large spread in the magnitude of this increase
17 reflecting structural uncertainties associated with the model representation of processes that influence
18 tropospheric ozone. Sources of uncertainties in SLCF abundance projections include scenario uncertainties,
19 or parametric and structural uncertainties in the model representation of the processes affecting simulated
20 abundances with implications for radiative forcing and air quality. The evolution of methane abundances in
21 SSP scenarios, for example, is derived from integrated assessment models (IAMs) which do not include the
22 effects from biogeochemical feedbacks (e.g., climate-driven changes in wetland emissions) (Meinshausen et
23 al., 2020) introducing uncertainty.

24
25 In summary, in SSPs, in addition to the socio-economic development and climate mitigation policies shaping
26 the GHG emission trajectories, the SLCF emission trajectories are also steered by varying levels of air
27 pollution control originating from SSP narratives and independent from climate mitigation. Consequently,
28 SSPs span a wider range of SLCF emissions than considered in the RCPs, better covering the diversity of
29 future options in air pollution management and SLCF-induced climate effects (*high confidence*). In addition
30 to SSP-driven emissions, the future evolution of SLCFs abundance is also sensitive to chemical and
31 biogeochemical feedbacks involving SLCFs, particularly natural emissions, whose magnitude and sign are
32 poorly constrained.

33 34 35 6.7.1.2 Future evolution of surface ozone and PM concentrations

36
37 The projection of air quality relevant abundances (surface ozone and PM_{2.5}) under the SSP scenarios are
38 assessed here. Future changes in global and regional annual mean surface ozone and PM_{2.5} driven by the
39 evolution of emissions as well as climate change have been quantified by CMIP6 models analysed in
40 AerChemMIP (Allen et al., 2020; Turnock et al., 2020; Allen et al., 2021).

41
42
43 [START FIGURE 6.20 HERE]

44
45 **Figure 6.20: Projected changes in regional annual mean surface ozone (O₃) (ppb) from 2015 to 2100 in different**
46 **shared socio-economic pathways (SSPs).** Each panel represents values averaged over the corresponding
47 land area (except for Global) shown on the map in Figure 6.7. Solid colored lines and shading indicate the
48 multi-model mean and $\pm 1 \sigma$ across the available CMIP6 models (Turnock et al., 2020; Allen et al., 2021)
49 for each scenario. Changes are relative to annual mean values calculated over the period 2005-2014 from
50 the historical experiment as indicated in the top left of each regional panel along with $\pm 1 \sigma$. For each
51 model all available ensemble members are averaged before being used to calculate the multi-model mean.
52 Ozone changes are also displayed in the Interactive Atlas. Further details on data sources and processing
53 are available in the chapter data table (Table 6.SM.1).

54
55 [END FIGURE 6.20 HERE]

1 Surface ozone increases continuously until 2050 across most regions in SSP3-7.0 and SSP5-8.5, (Turnock et
2 al., 2020), particularly over Eastern Asia, Southern Asia, Middle East, Africa and South East Asia and
3 developing Pacific where this increase can reach and even exceed 5ppb for annual mean averaged over land
4 areas (Figure 6.20). After 2050, surface ozone concentrations decrease in SSP5-8.5, reaching levels below
5 there 2005-2014 levels in most regions but and level off or continue to increase under SSP3-7.0. This
6 increase in surface ozone in the SSP5-8.5 scenario occurs despite emissions decrease of several ozone
7 precursors because of is due to CH₄ emissions increase until about 2080 in the absence of climate mitigation.
8 Ozone decreases over all regions in response to strong emissions mitigation in SSP1-1.9 and SSP1-2.6
9 (Turnock et al., 2020), with decreases of 5 to 10 ppb as soon as 2030 in North America, Europe, Eurasia,
10 Eastern Asia, Middle East and Southern Asia in their annual means over land areas . In most regions surface
11 ozone is reduced slightly or remains near present day values in the middle of the road scenario, SSP2-4.5. In
12 2100, the largest differences in surface ozone changes across the scenarios occur for the Middle East,
13 Southern Asia and Eastern Asia with differences ranging up to 40 ppb between SSP3-7.0 and SSP1-1.9 at the
14 end of the century. Despite discrepancies in the magnitude of changes, especially over North America,
15 Europe, Eurasia, Eastern Asia and Southern Asia, the models are in high agreement regarding the signs of
16 the changes with are thus assessed as of *high confidence*.

17
18 The strong abatement of ozone precursor emissions (except those of CH₄) (SSP3-7.0-lowSLCF-highCH₄)
19 lead to a decrease of global average surface ozone by 15% (6 ppb) between 2015 and 2055 (Allen et al.,
20 2020) and ozone decreases in all regions except Southern Asia. However, this decrease is two times larger
21 when CH₄ emissions are abated simultaneously (SSP3-7.0-lowSLCF-lowCH₄), underlying the importance of
22 methane emission reduction as an important lever to reduce ozone pollution (*high confidence*) (see also
23 6.6.4).

24
25
26 **[START FIGURE 6.21 HERE]**

27
28 **Figure 6.21: Future changes in regional 5-year mean surface PM_{2.5} from 2015 to 2100 in different shared socio-**
29 **economic pathways (SSPs).** PM_{2.5} stands for micrograms per cubic meter of aerosols with diameter less
30 than 2.5 μm and is calculated by summing up individual aerosol mass components from each model as:
31 black carbon + particulate organic matter + sulphate + 0.25 * sea salt + 0.1 * dust. Since not all CMIP6
32 models reported nitrate aerosol, it is not included here. See Figure 6.20 for further details. PM_{2.5} changes
33 are also displayed in the Interactive Atlas. Further details on data sources and processing are available in
34 the chapter data table (Table 6.SM.1).

35
36 **[END FIGURE 6.21 HERE]**

37
38
39 A decrease in surface PM_{2.5} concentrations is estimated for SSP1-1.9, SSP1-2.6 and SSP2-4.5 (Turnock et
40 al., 2020) (Figure 6.21). A decrease in PM_{2.5}, is also projected in SSP5-8.5 which does not consider any
41 climate mitigation but has a strong air pollution control. The decrease is largest in the regions with the
42 highest 2005-2014 mean concentrations (Middle East, South Asia and East Asia). Under the SSP3-7.0
43 scenario, PM_{2.5} is predicted to increase or remain at near present-day values across Asia; regions where
44 present-day concentrations are currently the highest. There is large model spread over regions with large
45 natural aerosol sources, for example, in North Africa, where dust sources are important. The mitigation of
46 non-methane SLCFs in the SSP3-7.0-lowSLCF-highCH₄ scenario is predicted to reduce PM_{2.5} by 25% (in
47 2055, relative to the SSP3-7.0 scenario) over global land surface areas (Allen et al., 2020).

48
49 The magnitude of the annual mean change in surface ozone and PM_{2.5} for all the SSPs (accounting for both
50 emission and climate change) is greater than that expected from climate change in isolation (Turnock et al.,
51 2020). The uncertainty in the projections comes from how natural emissions will respond to climate change.
52 However, multiple lines of evidence (along with those from Sections 6.2.2, 6.5, and 6.7.1) provide *high*
53 *confidence* (compared to *medium* in AR5) that changes in emissions, and in particular human-induced
54 emissions, will drive future air pollution levels rather than physical climate change.

55
56 In summary, future air pollution levels are strongly driven by precursor emission trajectories in the SSPs

1 with substantial reductions in global surface ozone and PM (when air pollution and climate change are both
 2 strongly mitigated, e.g., SSP1-2.6) to no improvement and even degradation (when no climate change
 3 mitigation and only weak air pollution control are considered, SSP3-7.0) (*high confidence*). In the latter case,
 4 PM levels are estimated to increase until 2050 over large parts of Asia and surface ozone pollution worsens
 5 over all continental areas throughout the whole century (*high confidence*). In scenarios without climate
 6 change mitigation but with strong air pollution control (SSP5-8.5), high methane levels hamper the decline
 7 in global surface ozone in the near term and only PM levels decrease (*high confidence*).

10 6.7.2 Evolution of future climate in response to SLCF emissions

12 6.7.2.1 Effects of SLCFs on ERF and climate response

14 This section assesses how the different spatial and temporal evolution of SLCF emissions in the SSPs affects
 15 the future global and regional ERFs, and GSAT and precipitation responses. In CMIP6, only a very limited
 16 set of simulations (all based on the SSP3-7.0 scenario) have been carried out with coupled ESMs to
 17 specifically address the future role of SLCFs (Collins et al., 2017), see also Sections 4.3 and 4.4. Note that
 18 the ScenarioMIP simulations (Section 4.3) include the SLCF emissions (as shown in Figures 6.18 and 6.19),
 19 however they cannot be used to quantify the effect of individual forcers. Coupled ESMs can in principle be
 20 used for this through a series of sensitivity simulations (e.g. Allen et al., 2020, 2021), but the amount of
 21 computer time required has made this approach prohibitive across the full SSP range. Therefore, to quantify
 22 the contribution from emissions of individual forcers spanning the range of the SSP scenarios to GSAT
 23 response, the analysis is mainly based on estimates using a two-layer emulator configuration derived from
 24 the medians of MAGICC7 and FaIRv1.6.2 (see Section 1.5.3.4, Cross Chapter Box 7.1, and Supplementary
 25 Material 7.SM5.2). The contribution from SLCFs to changes in GSAT have been calculated based on the
 26 global mean ERF for the various components as assessed in section 7.3.5, using the two-layer emulator for
 27 the climate response (Cross Chapter Box 7.1 and Supplementary Material 7.SM5.2).

29 The projections of GSAT for a broad group of forcing agents (aerosols, methane, tropospheric ozone, and
 30 HFCs with lifetimes lower than 50 years) for the SSP scenarios shows how much of the future warming or
 31 cooling (relative to 2019) can be attributed to the SLCFs (Figure 6.22). Note that during the first two
 32 decades, some of these changes in GSAT are due to emissions before 2019, in particular for the longer lived
 33 SLCFs such as methane and HFCs (see also Figure 6.15). The scenarios SSP3-7.0-lowSLCF-highCH₄ and
 34 SSP3-7.0-lowSLCF-lowCH₄ are special cases of the SSP3-7.0 scenario with strong, but realistic, reductions
 35 in non-methane SLCFs and all SLCFs respectively (Gidden et al., 2019).

37 As discussed in sections 6.2, 6.3 and 6.4, there are uncertainties relating emissions of SLCFs to changes in
 38 abundance (see Box 6.2) and further to ERF, in particular for aerosols and tropospheric ozone. Furthermore
 39 there are uncertainties related to the climate sensitivity, i.e., the relation between ERF and change in GSAT.
 40 Uncertainties in the ERF are assessed in Chapter 7 and calibrated impulse response function includes also the
 41 assessed range (see Box 7.1). There are also uncertainties related to the efficacies of the different SLCFs and
 42 time scales for the response, in particular for regional emissions (Schwarber et al., 2019; Yang et al., 2019b)
 43 that cannot be accounted for with the simple models used here.

46 **[START FIGURE 6.22 HERE]**

48 **Figure 6.22: Time evolution of the effects of short-lived climate forcers (SLCFs) and hydrofluorocarbons**
 49 **(HFCs) on global surface air temperature (GSAT) across the WGI core set of Shared Socio-**
 50 **Economic Pathways (SSPs).** Effects of net aerosols, methane, tropospheric ozone and HFCs (with
 51 lifetimes <50years), relative to year 2019 and to year 1750. The GSAT changes are based on the assessed
 52 historic and future evolution of Effective Radiative Forcing (Section 7.3.5). The temperature responses to
 53 the ERFs are calculated with an impulse response function with an equilibrium climate sensitivity of
 54 3.0°C for a doubling of atmospheric CO₂ (feedback parameter of -1.31 W m⁻² °C⁻¹, see CC-Box 7.1).
 55 Uncertainties are 5-95% ranges. The vertical bars to the right in each panel show the uncertainties for the
 56 GSAT change between 2019 and 2100. Further details on data sources and processing are available in the

chapter data table (Table 6.SM.1).

[END FIGURE 6.22 HERE]

Historical emissions have been updated until 2019 (see Supplementary Material of Chapter 7, 7.SM.1.3.1) and used for ERF for calculating GSAT in Figure 6.22. Year 2019 has been chosen as the base year to be consistent with the attributed temperature changes since 1750 (Figure 7.8). The warming attributed to SLCFs (methane, ozone and aerosols) over the last decade (Figure 7.8) constitutes about 30% of the peak SLCF driven warming in the most stringent scenarios (SSP1), in good agreement with Shindell and Smith (2019), and supported by the recent observed decline in AOD (Section 2.2.6).

From 2019 and until about 2040, SLCFs and HFCs will contribute to increase GSAT in the WG1 core set of SSP scenarios, with a *very likely* range of 0.04-0.41°C relative to 2019. The warming is most pronounced in the strong mitigation scenarios (i.e., SSP1-1.9 and SSP1-2.6) due to rapid cuts in aerosols. In the scenario SSP3-7.0, there is no reduction of aerosols until mid-century and it is the increases in methane and ozone that give a net warming in 2040. The warming is similar in magnitude to the SSP1-scenarios, in which the reduction in aerosols is the main driver. Contributions to warming from methane, ozone, aerosols and HFCs make SSP5-8.5 the scenario with the highest warming in 2040 and throughout the century.

After about 2040, it is *likely* that across the scenarios the net effect of the removal of aerosols is a further increase in GSAT. However, their contribution to the rate of change decreases towards the end of the century (from up to 0.2°C per decade before 2040 to about 0.03 °C per decade). After 2040, the changes in methane, HFCs and tropospheric ozone become equally important as the changes in the aerosols for the GSAT trends. In the low emission scenarios (SSP1-1.9 and SSP1-2.6), the contribution to warming from the SLCFs peaks around 2040 with a *very likely* range of 0.04°C-0.34°C. After the peak, the reduced warming from reductions in methane and ozone dominates, giving a best total estimate warming induced by SLCF and HFC changes of 0.12°C and 0.14°C respectively, in 2100, with a *very likely* range of -0.07°C to + 0.45°C (Figure 6.22). However, in the longer term towards the end of the century there are very significant differences between the scenarios. In SSP3-7.0 there is a near linear warming due to SLCFs of 0.08°C per decade, while for SSP5-8.5 there is a more rapid early warming. In SSP3-7.0, the limited reductions in aerosols, but a steady increase in methane, HFCs and ozone lead to a nearly linear contribution to GSAT reaching a best estimate of 0.5°C in 2100. Contributions from methane and ozone decrease towards 2100 in SSP5-8.5, however the warming from HFCs still increase and the SSP5-8.5 has the largest SLCF and HFC warming in 2100 with a best estimate of 0.6°C. In the SSP2-4.5 scenario, a reduction in aerosols contributes to about 0.3°C warming in 2100, while contributions from ozone and methane in this scenario are small.

The simplified approach used to estimate the contributions to GSAT in Figure 6.22 have been supplemented with ESM simulations driven by the two versions of the SSP3-7.0-lowSLCF scenario (Section 6.7.1.1). Results from five CMIP6 ESMs with fully interactive atmospheric chemistry and aerosols for the high-methane scenario show (Allen et al., 2020, 2021) that these reductions in emissions of air pollutants would lead to additional increase in GSAT by 2055 relative to 2015 compared to the standard SSP3-7.0 scenario, with a best estimate of $0.23 \pm 0.05^\circ\text{C}$, and a corresponding increase in global mean precipitation of $1.3 \pm 0.17\%$ (note that uncertainties from the work of Allen et al here and elsewhere are reported as twice standard deviation). Including methane mitigation (SSP3-7.0-lowSLCF-lowCH₄) would lead to a small increase in global precipitation ($0.7 \pm 0.1\%$) by mid-century despite a decrease in GSAT (see 6.7.3), which is related to the higher sensitivity of precipitation to sulphate aerosols than greenhouse gases (Allen et al., 2021) (also see Section 8.2.1).

Regionally inhomogeneous ERFs can lead to regionally dependent responses (see section 6.4.3). Mitigation of non-methane SLCFs over the period 2015 to 2055 (SSP3-7.0-lowSLCF-highCH₄ versus SSP3-7.0) will lead to positive ERF over land regions (Allen et al., 2020). There are large regional differences in the ERF from no significant trend over Northern Africa to about $0.5 \text{ W m}^{-2} \text{ decade}^{-1}$ for South Asia. The differences are mainly driven by differences in the reductions of sulphate aerosols. There is not a strong correspondence between regional warming and the ERF trends. As expected, the sensitivity (temperature change per unit

1 ERF) increases towards higher latitudes due to climate feedbacks and teleconnections. Regionally, the
2 warming rates are higher over continental regions, with highest increase in temperatures for Central and
3 Northern Asia and the Arctic in 2055 relative to 2015. The models agree on increasing global mean trend in
4 precipitation due to SLCFs, however precipitation trends over land are more uncertain (Allen et al., 2020), in
5 agreement with the relationship between aerosol and precipitation trends assessed in Chapter 8.

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ESM estimates of future concentrations of various SLCFs vary considerably even when using the same
future emission scenarios which is related to sources of model structural uncertainty in the several physical,
chemical and natural emission model parameterisations. The general uncertainties in understanding and
representing chemical and physical processes governing the life cycle of SLCFs (see Box 6.1) necessarily
also applies to simulations of future concentrations and ERF. In addition, how the models are able to
simulate climate changes (i.e., circulation and precipitation) that affects the dispersion and removal of
SLCFs constitute a structural uncertainty in the models. Also SLCF related climate feedbacks (e.g., NO_x
from lightning or BVOCs from vegetation) (Section 6.4.5) add to the uncertainty.

In the near term (2035-2040), it is *unlikely* that differences in the socio-economic developments (as
embedded in the SSPs) can lead to a discernible difference in the net effect of changes of SLCFs on GSAT.
This is because the intermodel spread in the estimated net effect of SLCFs on GSAT is as large as difference
between the scenarios due to compensating effects of change in emissions leading to cooling and warming.
However, in the longer term there is *high confidence* that the net warming of the SLCFs will be lower in the
mitigation scenarios (SSP1-1.9 and SSP1-2.6 that include reductions in methane emissions) than in the high-
emission scenarios (SSP3-7.0 and SSP5-8.5).

6.7.2.2 Effect of regional emissions of SLCFs on GSAT

For SLCFs with lifetimes shorter than typical mixing times in the atmosphere (days to weeks), the effects on
secondary forcing agents (e.g. tropospheric ozone, sulphate and nitrate aerosols) depend on where and when
the emissions occur due to non-linear chemical and physical processes. Also, the effective radiative forcing
following a change in concentrations depends on the local conditions (see Sections 6.2, 6.3 and 6.4).

While the emulators used for GSAT projections shown in figure 6.22 do not take the regional perspective
into account, the set of simulations performed within the Hemispheric Transport of Air Pollutants phase two
(HTAP2) project (Galmarini et al., 2017) allows for this perspective. The results from the chemistry-
transport model OsloCTM3 taking part in in HTAP-2 have been used by Lund et al. (2020) to derive
regional specific absolute global warming potentials (AGTPs, cf. (Aamaas et al., 2016)) for each emitted
SLCF and each HTAP-2 region. With this set of AGTPs Lund et al. (2020) estimate the transient response in
GSAT to the regional anthropogenic emissions. There are important differences in the contributions to
GSAT in 2040 and 2100 (relative to 2020) between the regions and scenarios mainly due to the differences
in mixture of emitted SLCFs (Figure 6.23). There is overall good agreement between the total net
contribution from all regions to GSAT and the estimate based on global ERF and the two-layer emulator
(Figure 6.22).

In the low and medium emission scenarios (SSP1-2.6 and SSP2-4.5), the warming effects of the SLCFs on
GSAT are dominated by emissions in North America, Europe, and East Asia (Figure 6.23). In SSP1-2.6 the
emissions of all SLCFs in all regions decrease and the net effect of the changes in SLCFs from all of these
three regions is an increase in GSAT of about 0.02°C (per region) in 2040 and about 0.04°C in 2100. For
SSP2-4.5 emissions of most SLCFs continue to increase in South Asia (see Figure 6.19), leading to a net
cooling in the near term (-0.03°C in 2040), while in 2100, North America, Europe, East and South Asia all
contribute to a warming, most pronounced from East Asia (0.05°C). In the SSP3-7.0 scenario the net effect
of SLCFs in all regions is an enhanced warming towards the end of the century. Methane then becomes the
dominant SLCF, and Africa is the region contributing the most to predicted global warming in 2100
(0.24°C). In SSP5-8.5, methane emissions increase in North America, Europe and Africa, while there is a
decrease in the Asian regions. For North America and Europe, the methane increase combined with a
reduction in aerosol leads to highest net contribution to GSAT in this scenario (0.06 and 0.04°C in 2100,
respectively). The high growth in methane makes Africa the region with the largest contribution to future

1 warming by SLCFs (0.18°C in 2100 versus 2020) in this scenario.
2
3

4 **[START FIGURE 6.23 HERE]**
5

6 **Figure 6.23: Contribution from regional emissions of short-lived climate forcers (SLCFs) to changes in global surface air temperature (GSAT) in 2040 (upper row) and 2100 (lower row), relative to 2020 for four Shared Socio-economic Pathways (SSP).** Adapted from (Lund et al., 2020). NO_x, CO, and VOC account for the impact through changes in ozone and methane, NO_x additionally includes the impact through formation of nitrate aerosols. BC, SO₂ and OC accounts for the direct aerosol effect (aerosol-radiation interactions), as well as an estimate of the semi-direct effect for BC due to rapid adjustments and indirect effect (aerosol-cloud interactions) of sulfate aerosols. Regions are the same as shown in the map in Figure 6.7. Further details on data sources and processing are available in the chapter data table (Table 6.SM.1).
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16 **[END FIGURE 6.23 HERE]**
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18

19 6.7.3 Effect of SLCFs mitigation in SSP scenarios 20

21 Air quality policies lead to a decrease in emissions of both warming and cooling SLCFs. Here we assess the
22 contribution of SLCFs to the total warming (including also the long-lived climate forcers) in the case of
23 stringent SLCF mitigation to improve air quality in scenarios with continued high use of fossil fuels (e.g.,
24 SSP3-7.0-lowSLCF and SSP5-8.5). Conversely, we also assess the effect on air quality of strategies aiming
25 to mitigate air pollution or climate change under the SSP3-7.0 framework (using the SSP3-7.0-lowSLCF-
26 lowCH₄, SSP3-7.0-lowSLCF-highCH₄ and SSP3-3.4 scenarios).
27

28 As illustrated in Figure 6.24 (Section 6.7.2.2) and in Figure 2.2 of SR1.5 (Rogelj et al., 2018a), the total
29 aerosol ERF change in stringent mitigation pathways is expected to be dominated by the effects from the
30 phase-out of SO₂ contributing to a warming. Recent emission inventories and observations of trends in AOD
31 (see Section 2.2.6 and 6.2.1) show that it is *very likely* that there have been reductions in global SO₂
32 emissions and in aerosol burdens over the last decade. Here, we use 2019 as the reference year rather than
33 the ‘Recent Past’ defined as the average over 1995–2014 (Section 4.1) in order to exclude the recent
34 emission reductions when discussing the different possible futures.
35

36 The role of the different SLCFs, and also the net of all the SLCFs relative to the total warming in the
37 scenarios, is quite different across the SSP scenarios varying with the summed levels of climate change
38 mitigation and air pollution control (Figure 6.24). In the scenario without climate change mitigation but with
39 strong air pollution control (SSP5-8.5) all the SLCFs (CH₄, aerosols, tropospheric O₃) and the HFCs (with
40 lifetimes less than 50 years) add to the warming, while in the strong climate change and air pollution
41 mitigation scenarios (SSP1-1.9 and SSP1-2.6), the emission controls act to reduce methane, ozone and BC,
42 and these reductions thus contribute to cooling. In all scenarios, except SSP3-7.0, emission controls lead to a
43 reduction of the aerosols relative to 2019, causing a warming. However, the warming from aerosol
44 reductions is stronger in the SSP1 scenarios (with best estimates of 0.21°C in 2040 and 0.4°C in 2100 in
45 SSP1-2.6) because of higher emission reductions from stronger decrease of fossil-fuel use in these scenario
46 than in SSP5-8.5 (0.13°C in 2040 and 0.22°C in 2100). The changes in methane abundance contribute a
47 warming of 0.14°C in SSP5-8.5, but a cooling of 0.14°C in SSP1-2.6 by the end of the 21st century relative
48 to 2019. Furthermore, under SSP5-8.5, HFCs induce a warming of 0.06°C with a *very likely* range of [0.04 to
49 0.08°C] in 2050 and 0.2 [0.1 to 0.3]°C by the end of the 21st century, relative to 2019, while under SSP1-
50 2.6, warming due to HFCs is negligible (below 0.02°C) (*high confidence*). This assessment relies on these
51 estimates, which are based on updated ERFs and HFC lifetimes. It is in accordance with previous estimates
52 (Section 6.6.3.2) of the efficiency of implementation of the Kigali Amendment and national regulations. It is
53 *very likely* that under a stringent climate and air pollution mitigation scenario (SSP1-2.6), the warming
54 induced by changes in methane, ozone, aerosols, and HFCs towards the end of 21st century, will be very low
55 compared with the warming they would cause under SSP5-8.5 scenario (0.14 °C in SSP1-2.6 versus 0.62 °C
56 in SSP5-8.5).

1
2 For the SSP3-7.0-lowSLCF-highCH₄ and SSP3-7.0-lowSLCF-lowCH₄ scenarios, a five ESMS ensemble has
3 been analyzed relative to the standard SSP3-7.0 scenario (Allen et al., 2020, 2021). For SSP3-7.0-lowSLCF-
4 highCH₄, in which the methane emissions are as in the standard SSP3-7.0 scenario, (Allen et al., 2021) found
5 an enhanced global mean surface warming of 0.23 ± 0.05 °C by mid-century and 0.21 ± 0.03 °C by 2100
6 relative to the warming in the standard SSP3-7.0 scenario. Including also strong mitigation of methane
7 emissions, the same models (Allen et al., 2021) find that the warming is off-set resulting in a net cooling of
8 0.15 ± 0.05 °C at mid-century (2050-2059) and 0.50 ± 0.02 °C at the end of the century (2090-2099) relative to
9 SSP3-7.0.

10
11 There is *robust evidence* and *high agreement* that non-methane SLCFs mitigation measures, through
12 reductions in aerosols and non-methane ozone precursors to improve air quality (SSP3-7.0-lowSLCF-
13 highCH₄ versus SSP3-7.0), would lead to additional near-term warming with a range of 0.1°C -0.3°C.
14 Methane mitigation which also reduces tropospheric ozone, stands out as an option that combines near and
15 long term gains on surface temperature (*high confidence*). With stringent but realistic methane mitigation
16 (SSP3-7.0-lowSLCF-lowCH₄), it is *very likely* that warming (relative to SSP3-7.0) from non-methane SLCFs
17 can be off-set (Figure 6.24 and Allen et al., 2021). Due to the slower response to the methane mitigation, this
18 offset becomes more robust over time and it is *very likely* to be an offset after 2050. However, when
19 comparing to present day, it is *unlikely* that that methane mitigation can fully cancel out the warming over
20 the 21st century from reduction of non-methane cooling SLCFs.

21
22 The SSP3 storyline assumes ‘regional rivalry’ (see 1.6.1.1) with weak air pollution legislation and no climate
23 mitigation and is compared here against SSP3-7.0-lowSLCF-lowCH₄ (strong air pollution control) and
24 SSP3-3.4 (the most ambitious climate policy feasible under SSP3 narrative). In the SSP3-3.4 scenario, all
25 emissions follow the SSP3-7.0 scenario until about 2030 and then deep and rapid cuts in fossil fuel use are
26 imposed (Fujimori et al., 2017). In the case of climate mitigation, such as in the SSP3-3.4 scenario, the
27 decrease of SLCFs emissions is a co-benefit from the targeted decrease of CO₂ (when SLCFs are co-
28 emitted), but also directly targeted as in the case of methane. For SLCFs, this means that emissions of
29 aerosols and methane increase until 2030 and are reduced fast thereafter (Fujimori et al., 2017). The effect on
30 GSAT (relative to 2019) is shown in Figure 6.22 and 6.24. The net GSAT response to the SLCFs is
31 dominated by the aerosols, with an initial cooling until 2030, then a fast rebound for 15 years followed by a
32 very moderate warming reaching 0.21°C in 2100. The ozone change causes a slight cooling (up to 0.06°C),
33 in contrast to the warming in the SSP3-7.0-lowSLCF-highCH₄ scenario in which the methane emissions
34 increase until 2100.

35
36
37 **[START FIGURE 6.24 HERE]**

38
39 **Figure 6.24: Effects of short-lived climate forcers (SLCFs) and hydrofluorocarbons (HFCs) on global surface air**
40 **temperature (GSAT) across the WGI core set of Shared Socio-Economic Pathways (SSPs).** Effects
41 of net aerosols, methane, tropospheric ozone, and hydrofluorocarbons (HFCs; with lifetimes < 50 years),
42 are compared with those of total anthropogenic forcing for 2040 and 2100 relative to year 2019. The
43 GSAT changes are based on the assessed historic and future evolution of Effective Radiative Forcing
44 (Section 7.3.5). The temperature responses to the ERFs are calculated with an impulse response function
45 with an equilibrium climate sensitivity of 3.0°C for a doubling of atmospheric CO₂ (feedback parameter
46 of $-1.31 \text{ W m}^{-2} \text{ °C}^{-1}$, see Cross-Chapter Box 7.1). Uncertainties are 5-95% ranges. The scenario total (grey
47 bar) includes all anthropogenic forcings (long- and short-lived climate forcers, and land use changes)
48 whereas the white points and bars show the net effects of SLCFs and HFCs and their uncertainties.
49 Further details on data sources and processing are available in the chapter data table (Table 6.SM.1).

50
51 **[END FIGURE 6.24 HERE]**

52
53
54 To assess the effect of dedicated air quality versus climate policy on air quality, PM_{2.5} and ozone indicators
55 were estimated for three SSP3 scenarios by applying a widely used approach for the analysis of air quality
56 implications for given emission scenarios (Rao et al., 2017; Van Dingenen et al., 2018; Vandyck et al., 2018)

1 and whose sensitivity of surface concentrations to emission changes is comparable to ESMs ensemble (see
2 Supplementary Material 6.SM.5). The assessment shows that both strong air pollution control and strong
3 climate mitigation, implemented independently, lead to large reduction of exposure to PM_{2.5} and ozone by
4 the end of the century (Figures 6.25, 6.26) (*high confidence*). However, implementation of air pollution
5 control, relying on the deployment of existing technologies, leads to benefits more rapidly than climate
6 mitigation (*high confidence*), which requires systemic changes and is thus implemented later in this scenario.
7 Notably, under the underlying SSP3 context, significant parts of the population remain exposed to air quality
8 exceeding the WHO guidelines for PM_{2.5} over the whole century (*high confidence*), in particular in Africa,
9 Eastern and Southern Asia, Middle East and for ozone only small improvement in population exposure is
10 expected in Africa and Asia. Confidence levels here result from expert judgement on the whole chain of
11 evidence.

12
13
14 **[START FIGURE 6.25 HERE]**

15
16 **Figure 6.25: Effect of dedicated air pollution or climate policy on population-weighted PM_{2.5} concentrations (µg**
17 **m⁻³) and share of population (%) exposed to different PM_{2.5} levels across selected world regions.**
18 Thresholds of 10 µg m⁻³ and 35 µg m⁻³ represent the WHO air quality guideline and the WHO interim
19 target 1, respectively; WHO (2017). Results are compared for SSP3-7.0 (no major improvement of
20 current legislation is assumed), SSP3-LowSLCF (strong air pollution controls are assumed), and a climate
21 mitigation scenario SSP3-3.4; details of scenario assumptions are discussed in Riahi et al. (2017) and Rao
22 et al. (2017). Analysis performed with the TM5-FASST model (Van Dingenen et al., 2018) using
23 emission projections from the Shared Socio-Economic Pathway (SSP) database (Riahi et al., 2017;
24 Rogelj et al., 2018a; Gidden et al., 2019). Further details on data sources and processing are available in
25 the chapter data table (Table 6.SM.1).

26
27
28 **[END FIGURE 6.25 HERE]**

29
30
31 **[START FIGURE 6.26 HERE]**

32
33 **Figure 6.26: Effect of dedicated air pollution or climate policy on population-weighted ozone concentrations**
34 **(SOMO0; ppb) and share of population (%) exposed to chosen ozone levels across ten world**
35 **regions.** Results are compared for SSP3-7.0 (no major improvement of current legislation is assumed),
36 SSP3-Low NTCF (strong air pollution controls are assumed), and a climate mitigation scenario (SSP3-
37 3.4); details of scenario assumptions are discussed in Riahi et al. (2017) and Rao et al. (2017). Analysis
38 performed with the TM5-FASST model (Van Dingenen et al., 2018) using emission projections from the
39 Shared Socio-Economic Pathway (SSP) database (Riahi et al., 2017; Rogelj et al., 2018a; Gidden et al.,
40 2019). Further details on data sources and processing are available in the chapter data table (Table
41 6.SM.1).

42
43 **[END FIGURE 6.26 HERE]**

44
45
46 In summary, the warming induced by SLCF changes is stable after 2040 in the WG1 core set of SSP
47 scenarios associated with lower global air pollution as long as methane emissions are also mitigated, but the
48 overall warming induced by SLCF changes is higher in scenarios in which air quality continues to deteriorate
49 (caused by growing fossil fuel use and limited air pollution control) (*high confidence*). In the SSP3-7.0
50 context, applying an additional strong air pollution control resulting in reductions in anthropogenic aerosols
51 and non-methane ozone precursors would lead to an additional near-term global warming of 0.1°C with a
52 *very likely* range of [-0.05 to 0.25]°C (compared with SSP3-7.0 for the same period). A simultaneous
53 methane mitigation consistent with SSP1's stringent climate mitigation policy implemented in the SSP3
54 world, could entirely alleviate this warming and even lead to a cooling of 0.1°C with a *very likely* range of [-
55 0.1 to 0.20]°C (compared with SSP3-7.0 for the same period). Across the SSPs, the reduction of CH₄, ozone
56 precursors and HFCs can make a 0.2°C [0.1–0.4]°C difference on GSAT in 2040 and a 0.8°C [0.5–1.3]°C
57 difference at the end of the 21st century (Figure 6.24), which is substantial in the context of the 2015 Paris

1 Agreement. Sustained methane mitigation, wherever it occurs, stands out as an option that combines near
2 and long term gains on surface temperature (*high confidence*) and leads to air pollution benefit by reducing
3 globally the surface ozone level (*high confidence*).
4

5 Strong air pollution control as well as strong climate change mitigation, implemented independently, lead to
6 large reduction of the exposure to air pollution by the end of the century (*high confidence*). Implementation
7 of air pollution control, relying on the deployment of existing technologies, leads more rapidly to air quality
8 benefits than climate change mitigation which requires systemic changes but, in both cases, significant parts
9 of the population remain exposed to air pollution exceeding the WHO guidelines.
10

11 6.8 Perspectives

12 Ice-core analyses can now inform trends for more SLCFs over the last millennium (such as light NMVOCs
13 or CO) and more proxies are available to inform about past emissions. However, preindustrial levels of
14 SLCFs are still relatively poorly constrained. In addition, recent trends in abundances of the various types of
15 aerosols and of NMVOCs suffer from the scarcity of observation networks in various parts of the world, in
16 particular in the Southern Hemisphere. Such network development is necessary to record and understand the
17 evolution of atmospheric composition.
18

19 Assessment of the future air pollution changes at the urban level require the use of high resolution model to
20 properly account for non-linearities in chemistry, specific urban structures, local meteorology as well as
21 temporal and spatial variations in emissions and population exposure. To assess the relevance of air pollution
22 reduction policies, regional air quality models are necessary and are still not implemented in many
23 developing countries. To properly apply such models, the quality of spatialized emission inventories is
24 essential, but the production of accurate emission inventories remain a challenge for lots of rapidly growing
25 urban areas. The emission reporting now planned in the official mandate of the Task Force on National
26 Greenhouse Gas Inventories (TFI) can be a step in this direction if accompanied by efforts on emission
27 spatial distribution. An integrated modelling framework associating global and high-resolution chemistry-
28 transport models with shared protocols is missing to allow a systematic assessment of future changes on air
29 quality at this scale.
30

31 In parallel, opportunities of progress may emerge from big data acquisition. Big data and their mining can
32 inform practices related to emissions or can document pollution levels if associated with massive deployment
33 of low cost sensors through citizen science. New generation satellite data will as well give access to sub-km
34 scale air pollution observations.
35

36 A systematic emission modeling framework is needed to assess the LLGHG emission changes associated
37 with SLCFs reductions induced by air pollution control in the SSP framework. The SLCF mediated effects
38 of large-scale technology deployment to allow climate mitigation, such as hydrogen energy production,
39 carbon capture and storage though amine filters or changes in agricultural practices to limit GHG emissions
40 and/or produce bioenergy are also not considered in the emission scenarios.
41

42 Since AR5, the complexity of ESMs has increased to include many chemical and biogeochemical processes.
43 These last are necessary to quantify non-CO₂ biogeochemical feedbacks on the Earth System resulting from
44 climate driven changes in atmospheric chemistry and SLCF emissions from natural systems and, in turn,
45 impacts of SLCFs on biogeochemical cycles. Enhanced understanding of the biological, chemical and
46 physical processes based on experimental and observational work has facilitated advances in the ESMs.
47 However, assessment of non-CO₂ biogeochemical feedbacks and SLCF effects on land and aquatic
48 ecosystem productivity still remains challenging due to the multiple complex processes involved and
49 limitations in observational constraints to evaluate the skill of ESMs in realistically simulating the processes.
50 Advances will come from better understanding of the processes and mechanisms, in particular at the
51 component interfaces. The development of high resolution ESMs will facilitate their evaluation against high
52 resolution observations.
53
54
55

1 Frequently Asked Questions

3 FAQ 6.1: What are short-lived climate forcers and how do they affect the climate?

5 *Short-lived climate forcers (SLCFs) are compounds such as methane and sulphate aerosols that warm or*
6 *cool the Earth's climate over shorter time scales – from days to years – than greenhouse gases like carbon*
7 *dioxide, whose climatic effect lasts for decades, centuries or more. Because SLCFs do not remain in the*
8 *atmosphere for very long, their effects on the climate are different from one region to another and can*
9 *change rapidly in response to changes in SLCF emissions. As some SLCFs also negatively affect air quality,*
10 *measures to improve air quality have resulted in sharp reductions in emissions and concentrations of those*
11 *SLCFs in many regions over the few last decades.*

13 The SLCFs include gases as well as tiny particles called *aerosols*, and they can have a warming or cooling
14 effect on the climate (FAQ 6.1, Figure 1). Warming SLCFs are either greenhouse gases (e.g., ozone or
15 methane) or particles like black carbon (also known as soot), which warm the climate by absorbing energy
16 and are sometimes referred to as *short-lived climate pollutants*. Cooling SLCFs, on the other hand, are
17 mostly made of aerosol particles (e.g., sulphates, nitrates and organic aerosols) that cool down the climate by
18 reflecting away more incoming sunlight.

19
20 Some SLCFs do not directly affect the climate but produce climate-active compounds and are referred to as
21 precursors. SLCFs are emitted both naturally and as a result of human activities, such as agriculture or
22 extraction of fossil fuels. Many of the human sources, particularly those based on combustion, produce
23 SLCFs at the same time as carbon dioxide and other long-lived greenhouse gases. Emissions have increased
24 since the start of industrialization, and humans are now the dominant source for several SLCFs and SLCF
25 precursors, such as sulphur dioxide (which produces sulphates) and nitrogen oxides (which produce nitrates
26 and ozone), despite strong reductions over the last few decades in some regions from efforts to improve air
27 quality.

28
29 The climatic effect of a chemical compound in the atmosphere depends on two things: (1) how effective it is
30 at cooling or warming the climate (its *radiative efficiency*) and (2) how long it remains in the atmosphere (its
31 *lifetime*). Because they have high radiative efficiencies, SLCFs can have a strong effect on the climate even
32 though they have relatively short lifetimes of up to about two decades after emission. Today, there is a
33 balance between warming and cooling from SLCFs, but this can change in the future.

34
35 The short lifetime of SLCFs constrains their effects in both space and time. First, of all the SLCFs, methane
36 and the short-lived halocarbons persist the longest in the atmosphere: up to two decades (FAQ 6.1, Figure 1).
37 This is long enough to mix in the atmosphere and to spread globally. Most other SLCFs only remain in the
38 atmosphere for a few days to weeks, which is generally too short for mixing in the atmosphere, sometimes
39 even regionally. As a result, the SLCFs are unevenly distributed and their effects on the climate are more
40 regional than those of longer-lived gases. Second, rapid (but sustained) changes in emissions of SLCFs can
41 result in rapid climatic effects.

42
43
44 In addition to the direct warming and cooling effects, SLCFs have many other consequences for the climate
45 system and for air quality (see FAQ 6.2). For instance, deposition of black carbon on snow darkens its
46 surface, which subsequently absorbs more solar energy, leading to more melting and more warming.
47 Aerosols also modify the properties of clouds, which has indirect cooling effects on the climate and causes
48 changes in local rainfall (see FAQ 7.2). Climate models indicate that SLCFs have altered atmospheric
49 circulation on local and even hemispheric scales (e.g., monsoons) as well as regional precipitation. For
50 instance, recent observations show that regional weather is influenced by strong regional contrasts in the
51 evolution of aerosol concentrations, particularly over South and East Asia.

52
53 Although policies to limit climate change and discussions of the so-called *remaining carbon budgets*
54 primarily focus on carbon dioxide (see FAQ 5.4), SLCFs can significantly affect temperature changes. It is
55 therefore important to understand how SLCFs work and to quantify their effects. Because reducing some of

1 the SLCF emissions, such as methane, can simultaneously reduce warming effects and adverse effects on air
2 quality as well as help attaining Sustainable Development Goals, mitigation of SLCFs is often viewed as a
3 favourable ‘win-win’ policy option.
4
5

6 **[START FAQ 6.1, FIGURE 1 HERE]**
7

8 **FAQ 6.1, Figure 1: Main short-lived climate forcers, their sources, how long they exist in the atmosphere, and**
9 **their relative contribution to global surface temperature changes between 1750 and 2019**
10 (area of the globe). By definition, this contribution depends on the lifetime, the warming/cooling
11 potential (radiative efficiency), and the emissions of each compound in the atmosphere. Blue
12 indicates cooling and orange warming. Note that, between 1750 and 2019, the cooling contribution
13 from aerosols (blue diamonds and globe) was approximately half the warming contribution from
14 carbon dioxide.
15

16 **[END FAQ 6.1, FIGURE 1 HERE]**
17
18
19

20 **FAQ 6.2: What are the links between limiting climate change and improving air quality?**
21

22 *Climate change and air quality are intimately linked. Many of the human activities that produce long-lived*
23 *greenhouse gases also emit air pollutants, and many of these air pollutants are also ‘short-lived climate*
24 *forcers’ that affect the climate. Therefore, many options for improving air quality may also serve to limit*
25 *climate change and vice versa. However, some options for improving air quality cause additional climate*
26 *warming, and some actions that address climate change can worsen air quality.*
27

28 Climate change and air pollution are both critical environmental issues that are already affecting humanity.
29 In 2016, the World Health Organization attributed 4.2 million deaths worldwide every year to ambient
30 (outdoor) air pollution. Meanwhile, climate change impacts water resources, food production, human health,
31 extreme events, coastal erosion, wildfires, and many other phenomena.
32

33 Most human activities, including energy production, agriculture, transportation, industrial processes, waste
34 management and residential heating and cooling, result in emissions of gaseous and particulate pollutants
35 that modify the composition of the atmosphere, leading to degradation of air quality as well as to climate
36 change. These air pollutants are also *short-lived climate forcers* – substances that affect the climate but
37 remain in the atmosphere for shorter periods (days to decades) than long-lived greenhouse gases like carbon
38 dioxide (see FAQ 6.1). While this means that the issues of air pollution and climate change are intimately
39 connected, air pollutants and greenhouse gases are often defined, investigated and regulated independently of
40 one another in both the scientific and policy arenas.
41

42 Many sources simultaneously emit carbon dioxide and air pollutants. When we drive our fossil fuel vehicles
43 or light a fire in the fireplace, it is not just carbon dioxide or air pollutants that are emitted, but always both.
44 It is therefore not possible to separate emissions into two clearly distinct groups. As a result, policies aiming
45 at addressing climate change may have benefits or side-effects for air quality, and vice versa.
46

47 For example, some short-term ‘win-win’ policies that simultaneously improve air quality and limit climate
48 change include the implementation of energy efficiency measures, methane capture and recovery from solid
49 waste management and oil and gas industry, zero-emission vehicles, efficient and clean stoves for heating
50 and cooking, filtering of soot (particulate matter) for diesel vehicles, cleaner brick kiln technology, practices
51 that reduce burning of agricultural waste, and the eradication of burning of kerosene for lighting.
52

53 There are, however, also ‘win-lose’ actions. For example, wood burning is defined as carbon neutral because
54 a tree accumulates the same amount of carbon dioxide throughout its lifetime as is released when wood from
55 that tree is burned. However, burning wood can also result in significant emissions of air pollutants,
56 including carbon monoxide, nitrogen oxides, volatile organic compounds, and particulate matter, that locally

1 or regionally affect the climate, human health and ecosystems (FAQ 6.2, Figure 1). Alternatively, decreasing
2 the amount of sulphate aerosols produced by power and industrial plants and from maritime transport
3 improves air quality but results in a warming influence on the climate, because those sulphate aerosols
4 contribute to cooling the atmosphere by blocking incoming sunlight.
5

6 Air quality and climate change represent two sides of the same coin, and addressing both issues together
7 could lead to significant synergies and economic benefits while avoiding policy actions that mitigate one of
8 the two issues but worsen the other.
9

10
11 **[START FAQ 6.2, FIGURE 1 HERE]**
12

13 **FAQ 6.2, Figure 1: Links between actions aiming to limit climate change and actions to improve air quality.**

14 Greenhouse gases and aerosols (orange and blue) can affect directly climate. Air pollutants
15 (bottom) can affect the human health, ecosystems and climate. All these compounds have common
16 sources and sometimes interact with each other in the atmosphere which makes impossible to
17 consider them separately (dotted grey arrows).
18

19 **[END FAQ 6.2, FIGURE 1 HERE]**
20
21

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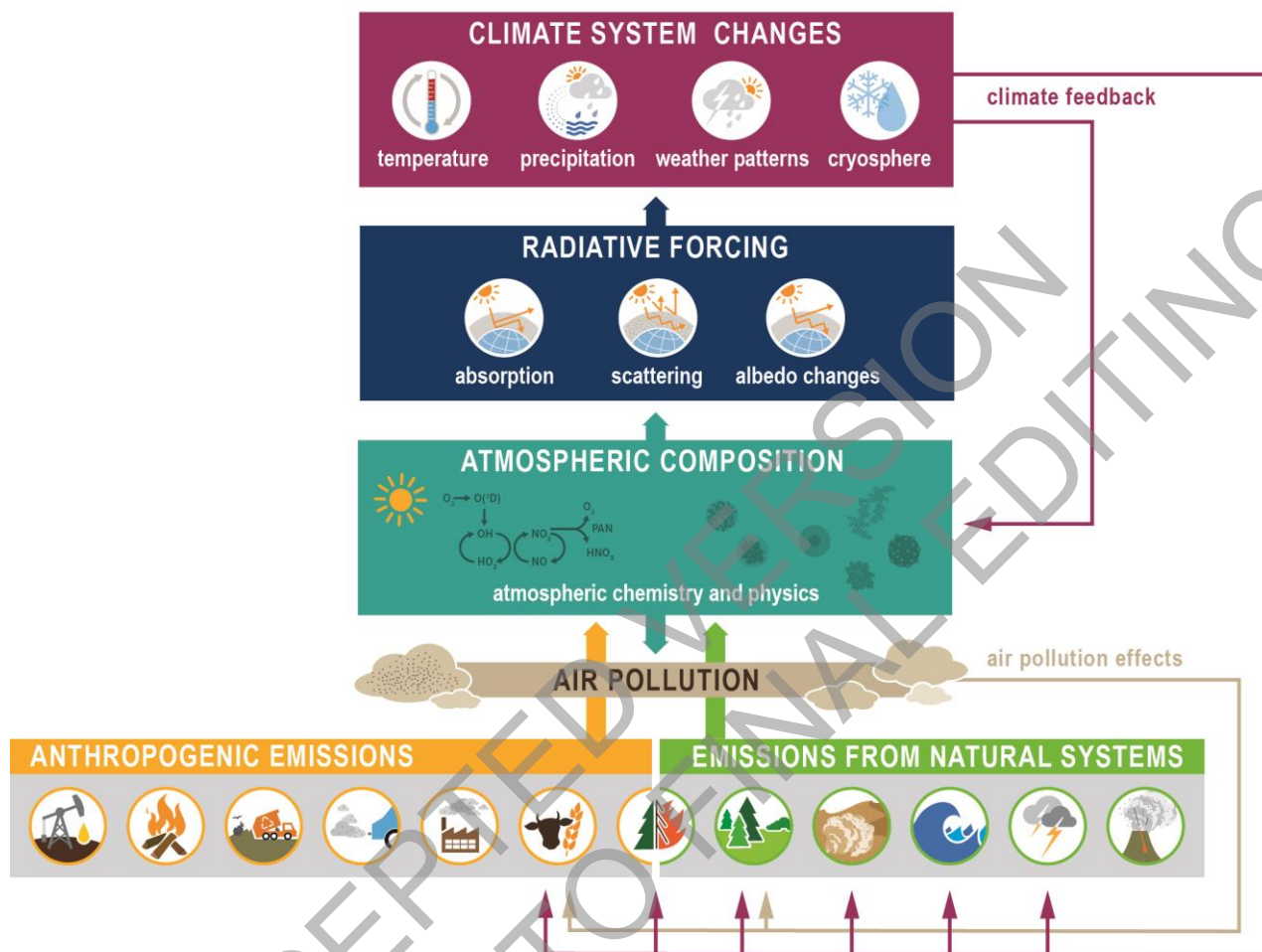
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Figures

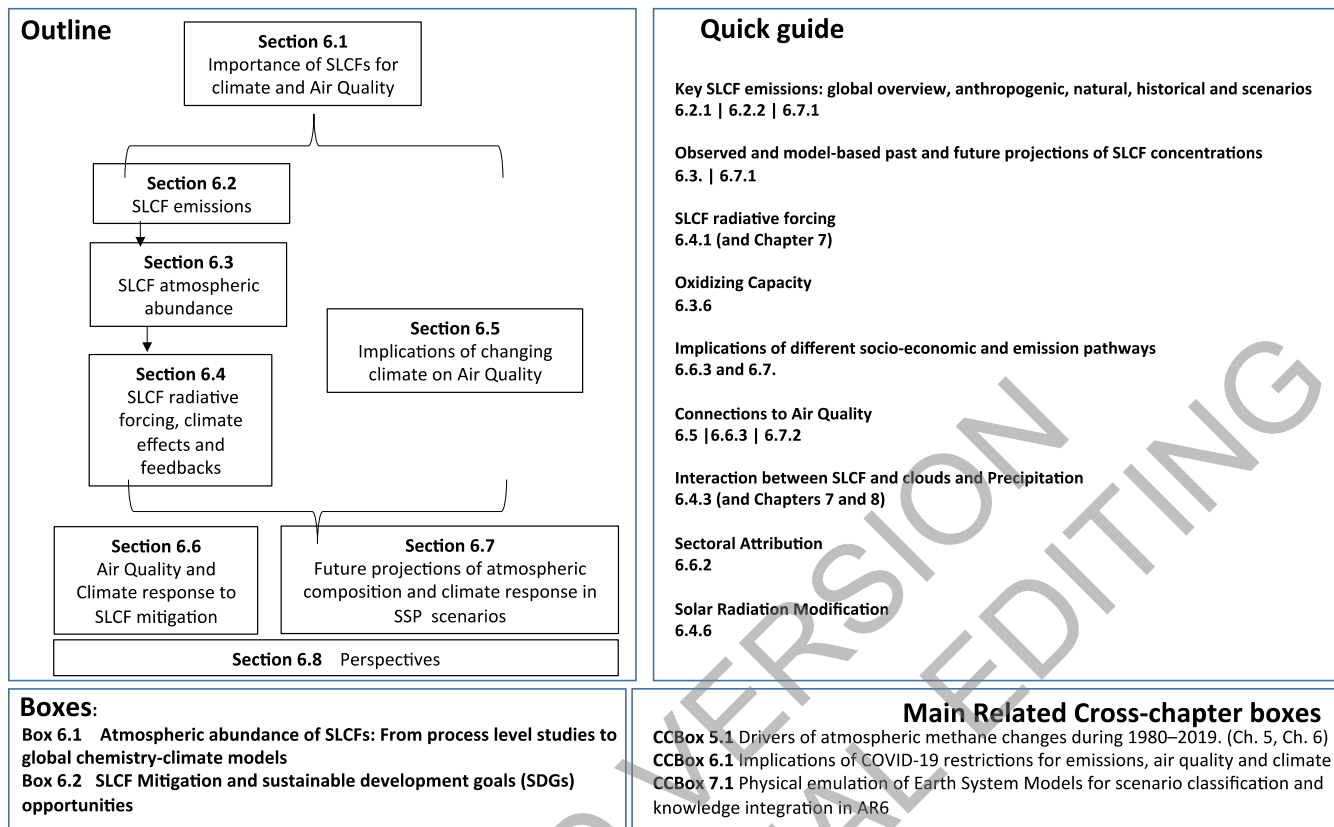
Sources and processes contributing to SLCFs and their effects on the climate system



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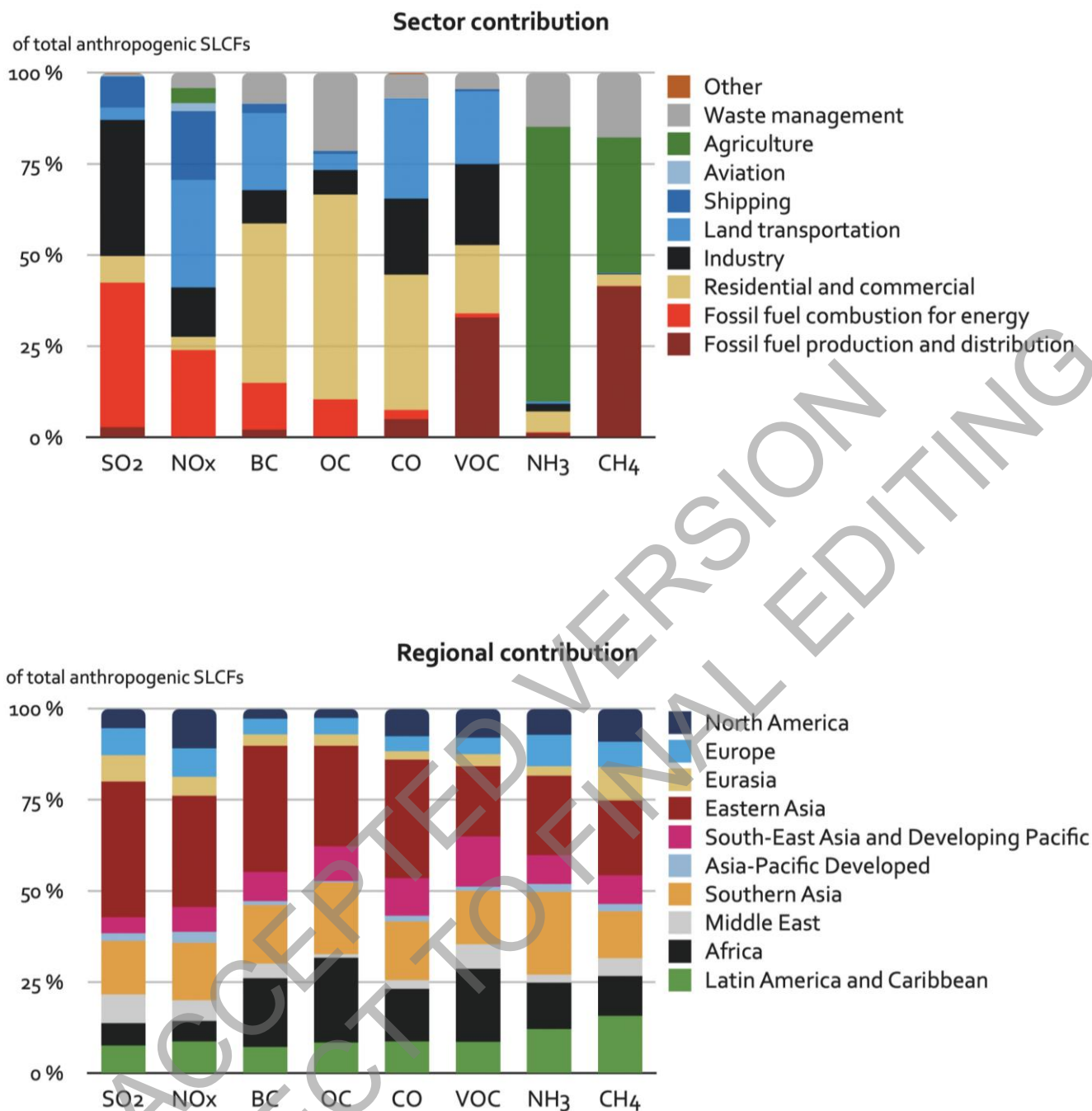
Figure 6.1: Sources and processes leading to atmospheric short-lived climate forcer (SLCF) burden and their interactions with the climate system. Both direct and indirect SLCFs and the role of atmospheric processes for the lifetime of SLCFs are depicted. Anthropogenic emission sectors illustrated are fossil fuel exploration, distribution and use, biofuel production and use, waste, transport, industry, agricultural sources, and open biomass burning. Emissions from natural systems include those from open biomass burning, vegetation, soil, oceans, lightning, and volcanoes. SLCFs interact with solar or terrestrial radiation, surface albedo, and cloud or precipitation system. The radiative forcing due to individual SLCF can be either positive or negative. Climate change induces changes in emissions from most natural systems as well as from some anthropogenic emission sectors (e.g. agriculture) leading to a climate feedback (purple arrows). Climate change also influences atmospheric chemistry processes, such as chemical reaction rates or via circulation changes, thus affecting atmospheric composition leading to a climate feedback. Air pollutants influence emissions from terrestrial vegetation, including agriculture (the grey arrow).

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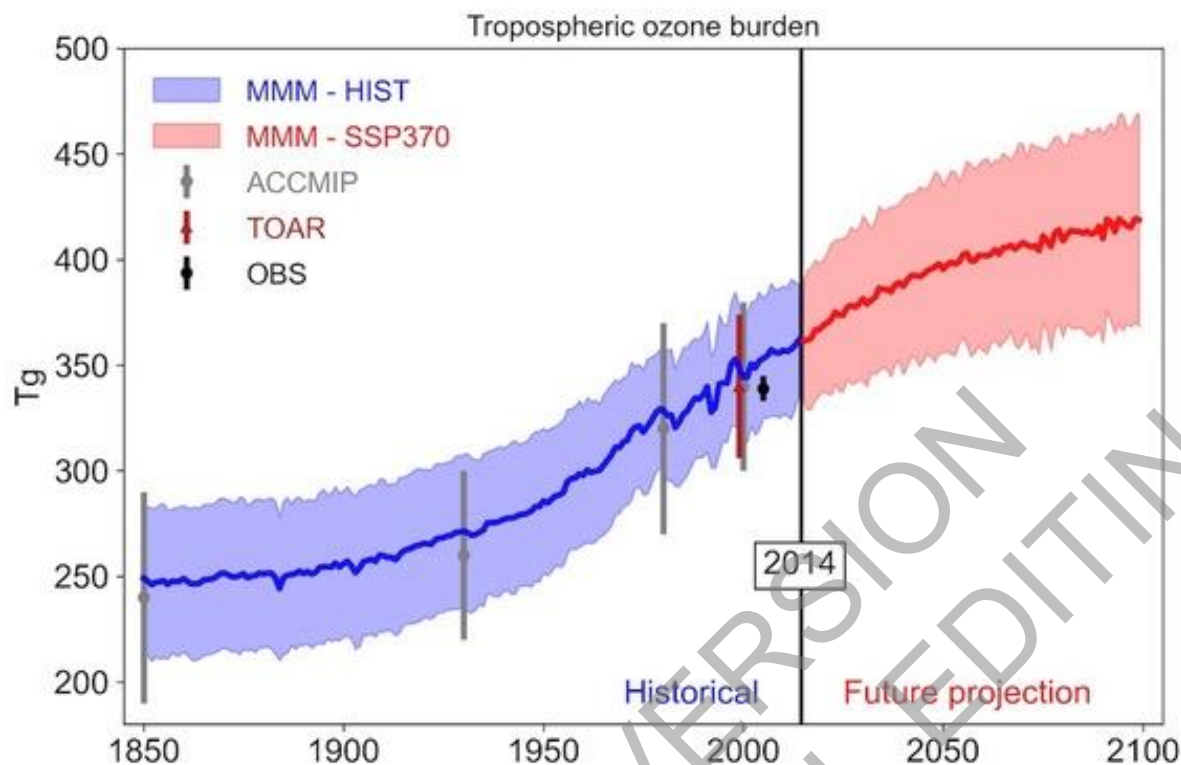
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Figure 6.2: Chapter roadmap. See Section 6.1.3 for additional description of the chapter.



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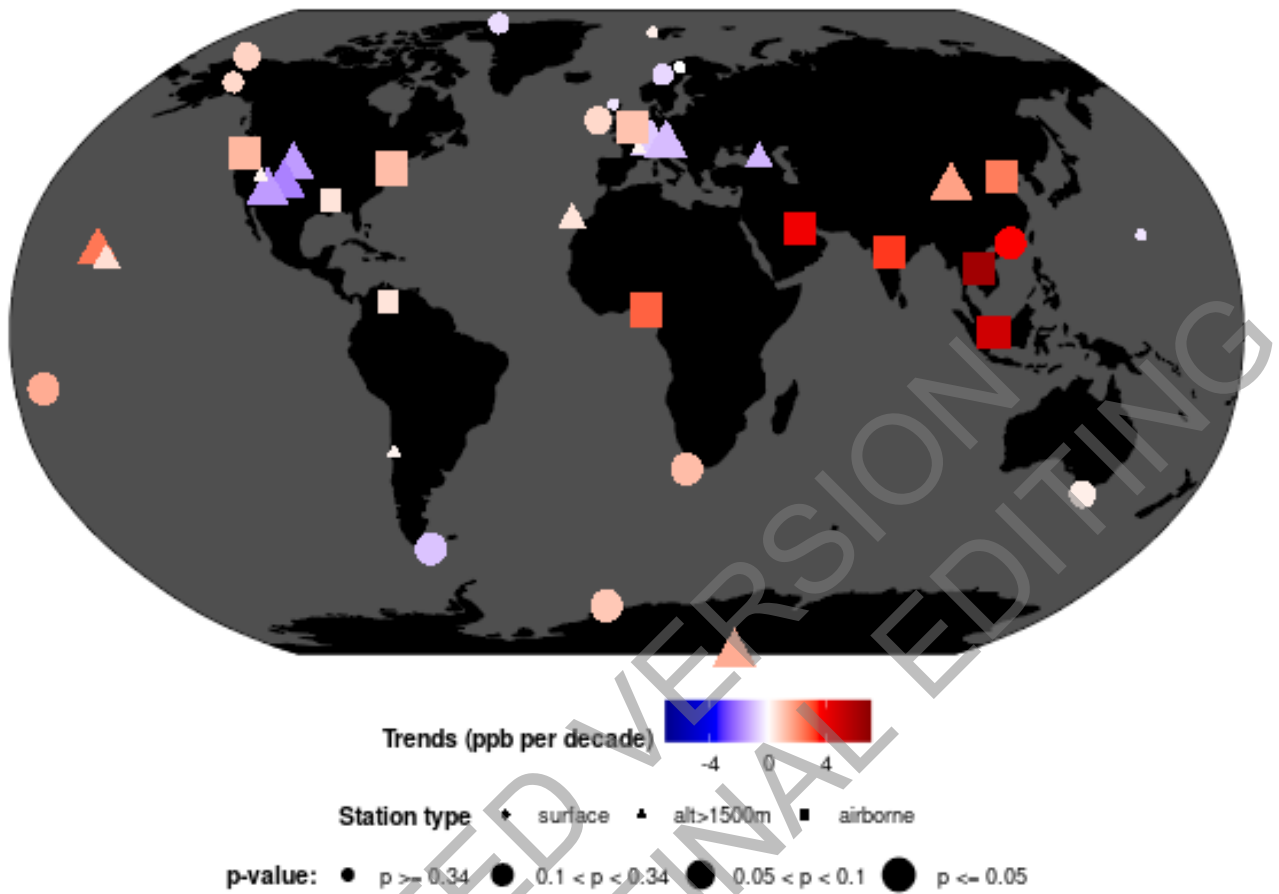
Figure 6.3: Relative regional and sectoral contributions to the present day (year 2014) anthropogenic emissions of Short Lived Climate Forcers (SLCFs). Emission data are from the Community Emissions Data System (CEDS) (Hoesly et al., 2018). Emissions are aggregated into the following sectors: fossil fuel production and distribution (coal mining, oil and gas production, upstream gas flaring, gas distribution networks), fossil fuel combustion for energy (power plants), residential and commercial (fossil and biofuel use for cooking and heating), industry (combustion and production processes, solvent use losses from production and end use), transport (road and off-road vehicles), shipping (including international shipping), aviation (including international aviation), agriculture (livestock and crop production), waste management (solid waste, including landfills and open trash burning, residential and industrial waste water), and other. Further details on data sources and processing are available in the chapter data table (Table 6.SM.1).



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2 **Figure 6.4:** Time evolution of global annual mean tropospheric ozone burden (in Tg) from 1850 to 2100. Multi-
3 model means for CMIP6 historical experiment (1850-2014) from UKESM1-LL-0, CESM2-WACCM,
4 MRI-ESM2-0, GISS-E2.1-G and GFDL-ESM4 and for ScenarioMIP SSP3-7.0 experiment (2015-2100)
5 are represented with their intermodel standard deviation ($\pm 1 \sigma$, shaded areas). Observation based global
6 tropospheric ozone burden estimate (from Table 6.3) is for 2010-2014. Tropospheric Ozone Assessment
7 Report (TOAR) multi-model mean value (from Table 6.3) is for 2000 with a $\pm 1 \sigma$ error-bar. Atmospheric
8 Chemistry and Climate Model Intercomparison Project (ACCMIP) multi-model means are for 1850,
9 1930, 1980 and 2000 time-slices with $\pm 1 \sigma$ error-bars. The troposphere is masked by the tropopause
10 pressure calculated in each model using the WMO thermal tropopause definition. Further details on data
11 sources and processing are available in the chapter data table (Table 6.SM.1).
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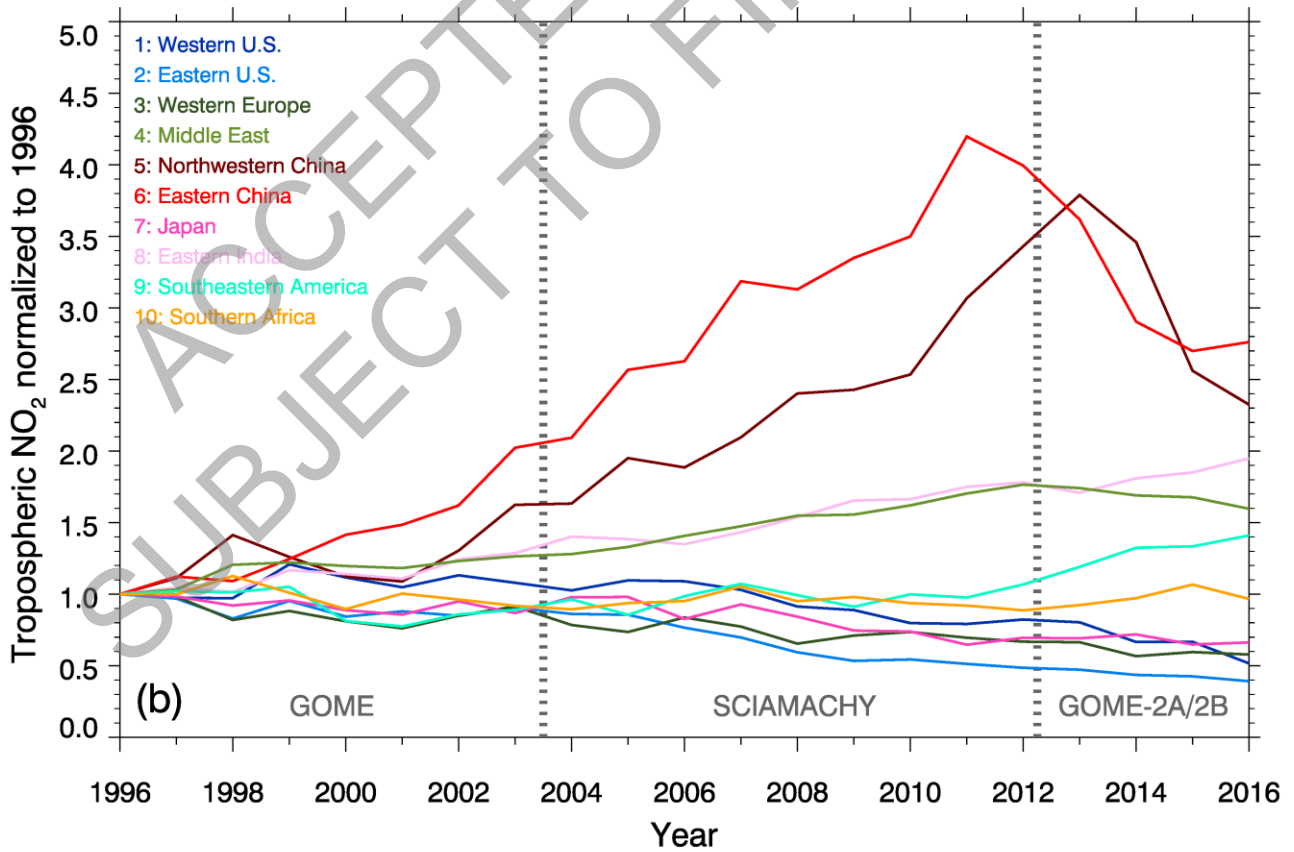
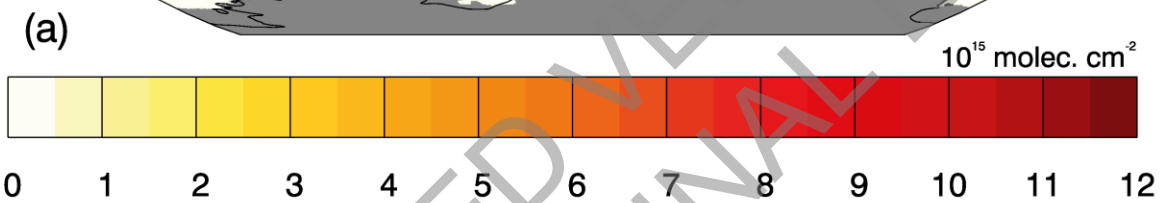
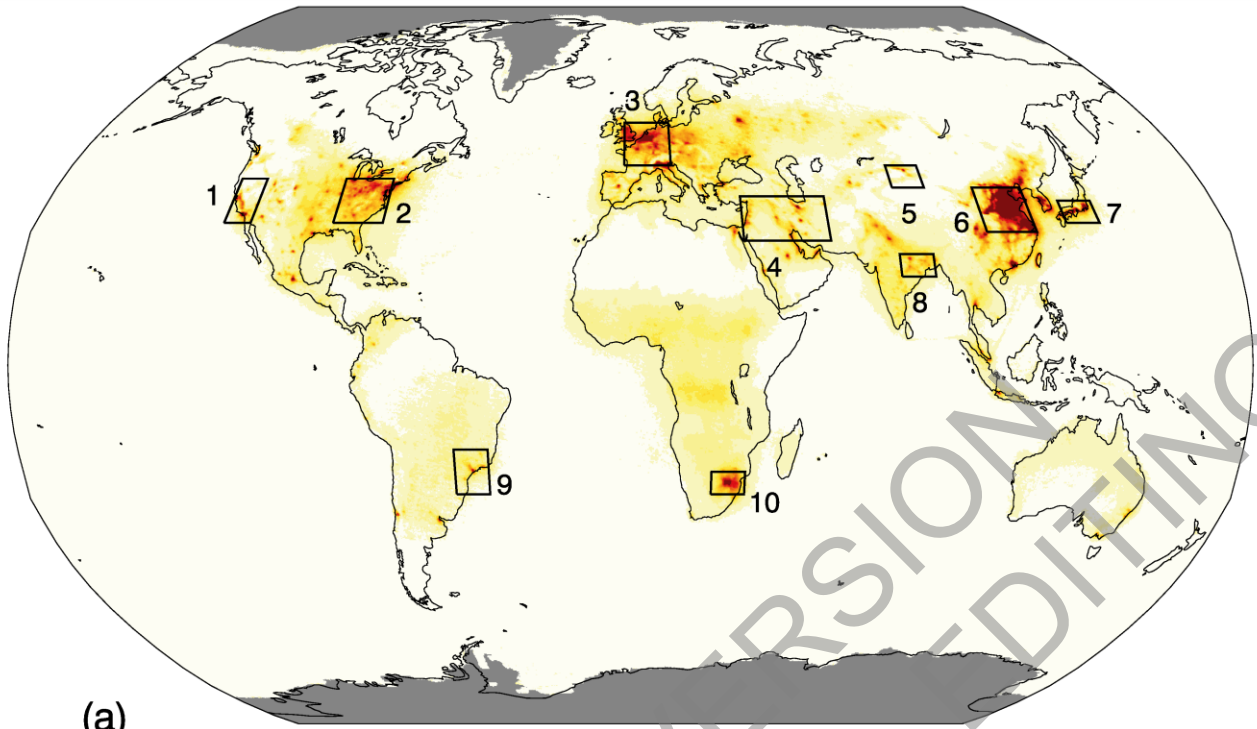
Decadal ozone trends since 1994



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Figure 6.5: Decadal tropospheric ozone trends since 1994. Trends are shown at 28 remote and regionally representative surface sites (Cooper et al., 2020) and in 11 regions of the lower free troposphere (650 hPa, ca. 3.5 km) as measured by In-Service Aircraft for a Global Observing System (IAGOS) above Europe, northeastern USA, southeastern USA, western North America, northeast China, South East Asia, southern India, Persian Gulf, Malaysia/Indonesia, Gulf of Guinea and northern South America (Gaudel et al., 2020). High elevation surface sites are >1500 m above sea level. All trends end with the most recently available year but begin in 1995 or 1994. The sites and datasets are the same as those used in Figure 2.8, further details on data sources and processing are available in the Chapter 2 data table (Table 2.SM.1).

Long term average and trends in tropospheric NO₂ column over 1996-2017



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Figure 6.6: Long term climatological mean (a) and time evolution (b) of tropospheric NO₂ vertical column density from the merged GOME/SCIAMACHY/GOME-2 (TM4NO2A version 2.3) dataset for the period 1996-2016 (Georgoulias et al., 2019). Time evolution of NO₂ column shown in panel (b) is normalized to the fitted 1996 levels for the 10 regions shown as boxes in panel (a). Further details on data sources and processing are available in the chapter data table (Table 6.SM.1).

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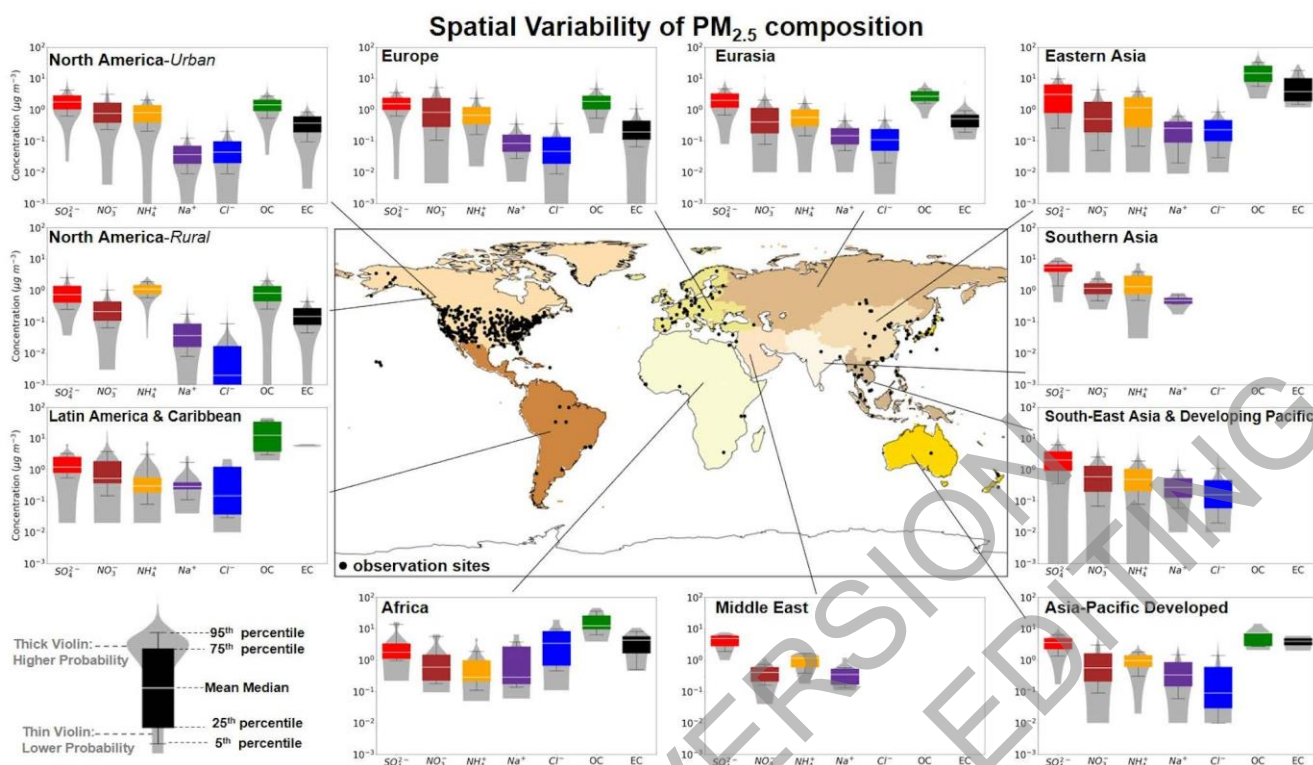
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Figure 6.7: Distribution of PM_{2.5} composition mass concentration (in $\mu\text{g m}^{-3}$) for the major PM_{2.5} aerosol components. Those aerosol components are sulfate, nitrate, ammonium, sodium, chloride, organic carbon, and elemental carbon. The central world map depicts the intermediate level regional breakdown of observations (10 regions) following the IPCC Sixth Assessment Report Working Group III (AR6 WGIII). Monthly averaged PM_{2.5} aerosol component measurements are from: (i) the Environmental Protection Agency (EPA) network which include 211 monitor sites primarily in urban areas of North America during 2000-2018 (Solomon et al., 2014) (ii) the Interagency Monitoring of Protected Visual Environments (IMPROVE) network during 2000-2018 over 198 monitoring sites representative of the regional haze conditions over North America, (iii) the European Monitoring and Evaluation Programme (EMEP) network over 70 monitoring in Europe and (eastern) Eurasia during 2000-2018, (iv) the Acid Deposition Monitoring Network in East Asia (EANET) network with 39 (18 remote, 10 rural, 11 urban) sites in Eurasia, Eastern Asia, South-East Asia and Developing Pacific, and Asia-Pacific Developed during 2001-2017, (v) the global Surface Particulate Matter Network (SPARTAN) during 2013-2019 with sites primarily in highly populated regions around the world (i.e., North America, Latin America and Caribbean, Africa, Middle East, Southern Asia, Eastern Asia, South-Eastern Asia and Developing Pacific) (Snider et al., 2015, 2016), and (vii) individual observational field campaign averages over Latin America and Caribbean, Africa, Europe, Eastern Asia, and Asia-Pacific Developed (Celis et al., 2004; Feng et al., 2006; Bourotte et al., 2007; Fuzzi et al., 2007; Mariani and de Mello, 2007; Molina et al., 2007; Favez et al., 2008; Mkoma, 2008; Aggarwal and Kawamura, 2009; Mkoma et al., 2009; de Souza et al., 2010; Li et al., 2010; Martin et al., 2010; Molina et al., 2010; Radhi et al., 2010; Weinstein et al., 2010; Batmunkh et al., 2011; Gioda et al., 2011; Pathak et al., 2011; Zhang et al., 2012; Cho and Park, 2013; Zhao et al., 2013; Wang et al., 2019; Kuzu et al., 2020). Further details on data sources and processing are available in the chapter data table (Table 6.SM.1).

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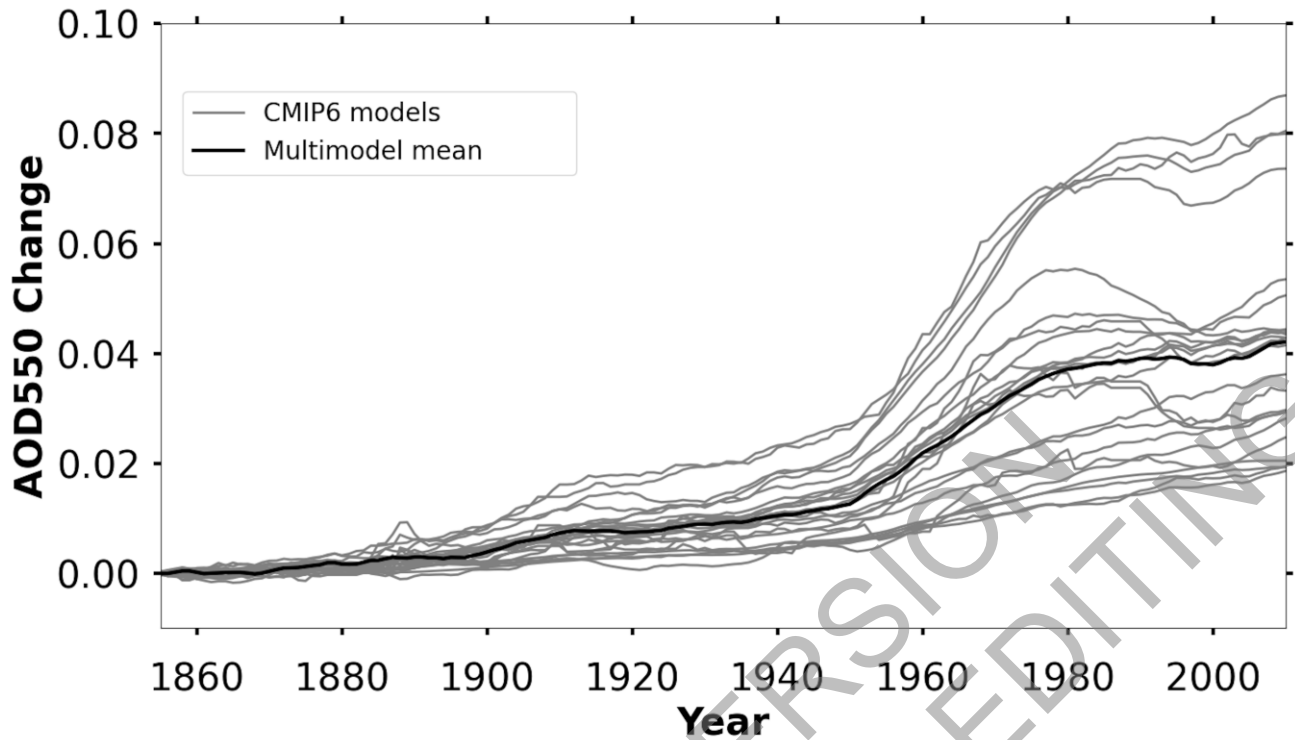
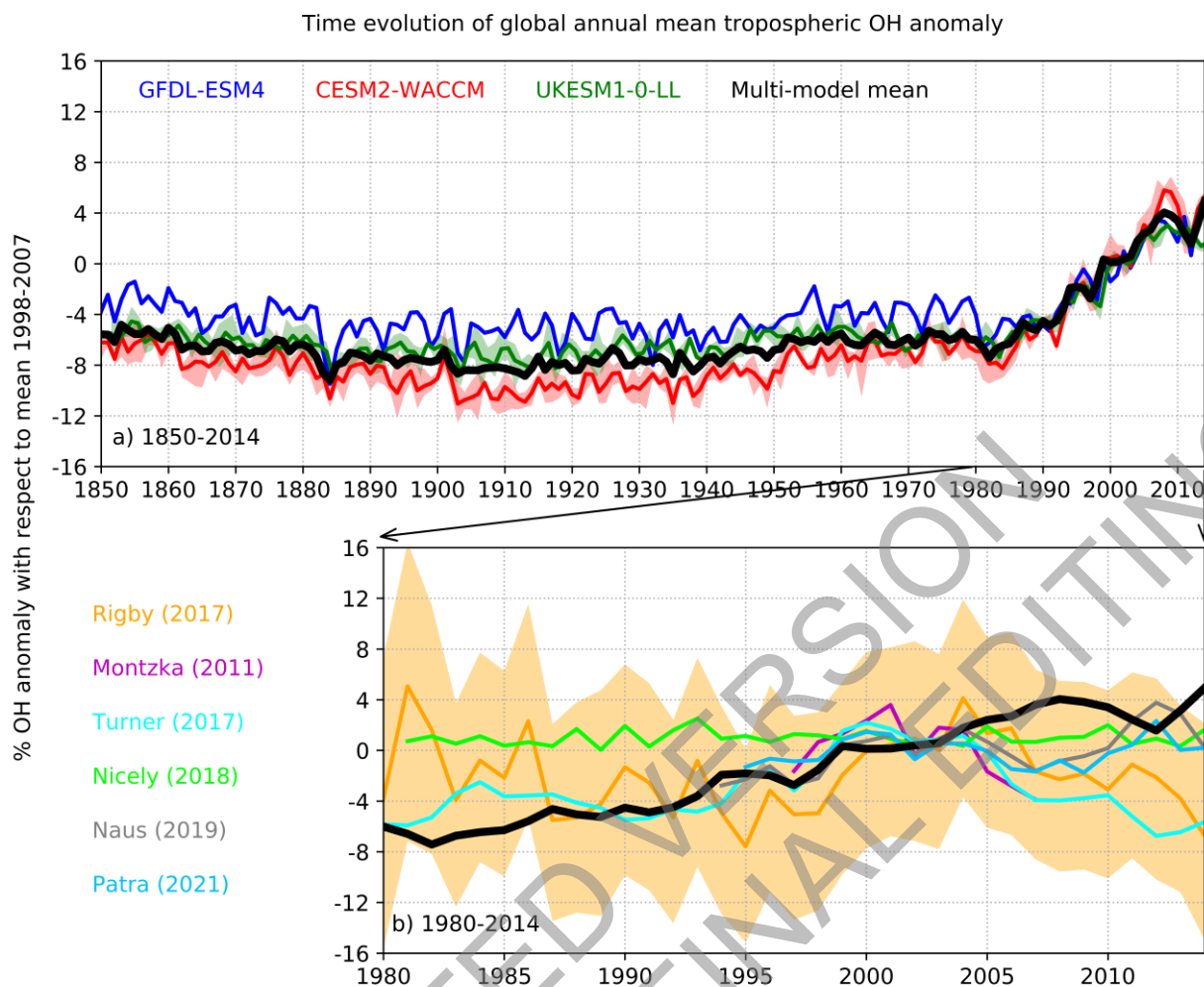
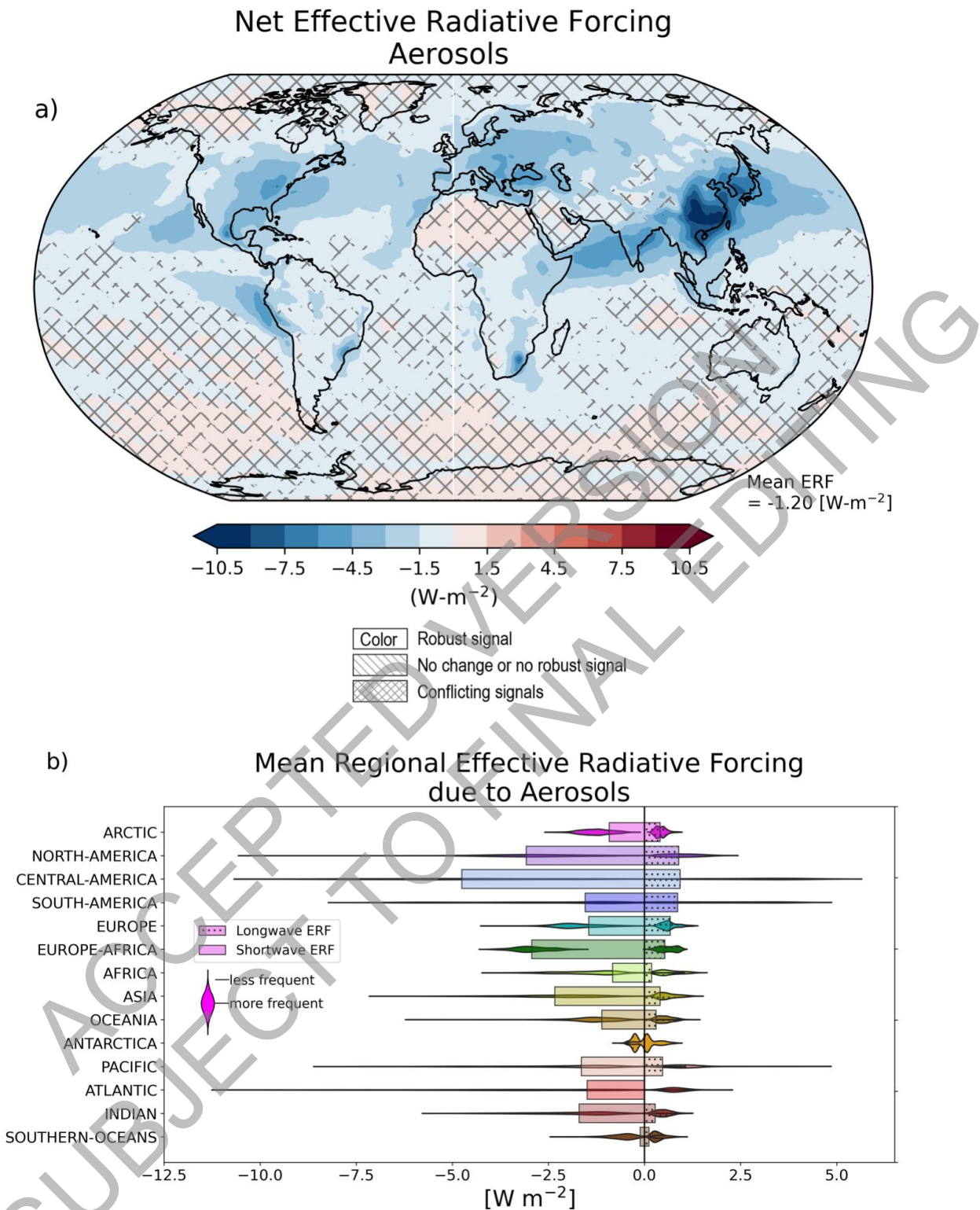
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Figure 6.8: Time evolution of changes in global mean aerosol optical depth (AOD) at 550 nm. Year of reference is 1850. Data shown from individual Coupled Model Intercomparison Project Phase 6 (CMIP6) historical simulations. Each time series corresponds to the ensemble mean of realizations done by each model. Simulation results from years including major volcanic eruptions, e.g. Novarupta (1912) and Pinatubo (1991), are excluded from the analysis for models encompassing the contribution of stratospheric volcanic aerosols to total AOD. Further details on data sources and processing are available in the chapter data table (Table 6.SM.1).



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Figure 6.9: Time evolution of global annual mean tropospheric hydroxyl (OH) over the historical period, expressed as a percentage anomaly relative to the mean over 1998-2007. a) Results from three CMIP6 models, including UKESM1-0LL (green), GFDL-ESM4 (blue), and CESM2-WACCM (red), are shown; the shaded light green and light red bands show mean over multiple ensemble members for UKESM1-0LL (3) and CESM2-WACCM (3) models, respectively with the multi-model mean anomalies shown in thick black line. b) multimodel mean OH anomalies for 1980-2015 period compared with those derived from observational-based inversions from (Montzka et al., 2011; Rigby et al., 2017; Turner et al., 2017; Nicely et al., 2018; Naus et al., 2019; Patra et al., 2021) in the zoomed box. Further details on data sources and processing are available in the chapter data table (Table 6.SM.1).



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Figure 6.10: Multi-model mean Effective radiative forcings (ERFs) due to aerosol changes between 1850 and recent-past (1995-2014). Panel (a) shows the spatial distribution of the net ERF with area-weighted global mean ERF shown at the lower right corner. Uncertainty is represented using the advanced approach: No overlay indicates regions with robust signal, where $\geq 66\%$ of models show change greater than variability threshold and $\geq 80\%$ of all models agree on sign of change; diagonal lines indicate regions with no change or no robust signal, where $< 66\%$ of models show a change greater than the variability threshold; crossed lines indicate regions with conflicting signal, where $\geq 66\%$ of models show change

1 greater than variability threshold and <80% of all models agree on sign of change. For more information
2 on the advanced approach, please refer to the Cross-Chapter Box Atlas.1. Panel (b) shows the mean
3 shortwave and longwave ERF for each of the 14 regions defined in the Atlas. Violins in panel (b) show
4 the distribution of values over regions where ERFs are significant. ERFs are derived from the difference
5 between top of the atmosphere (TOA) radiative fluxes for Aerosol Chemistry Model Intercomparison
6 Project (AerChemMIP) experiments *histSST* and *histSST-piAer* (Collins et al., 2017) averaged over 1995-
7 2014 (Box 1.4, Chapter 1). The results come from 7 Earth System Models: MIROC6, MPI-I-ESM-1-2-
8 HAM, MRI-ESM2-0, GFDL-ESM4, GISS-E2-1-G, NorESM2-LM, and UKESM-0-LL. These data can
9 be seen in the Interactive Atlas. Further details on data sources and processing are available in the chapter
10 data table (Table 6.SM.1).
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Temporal Regional Mean Net Effective Radiative Forcing due to Aerosols

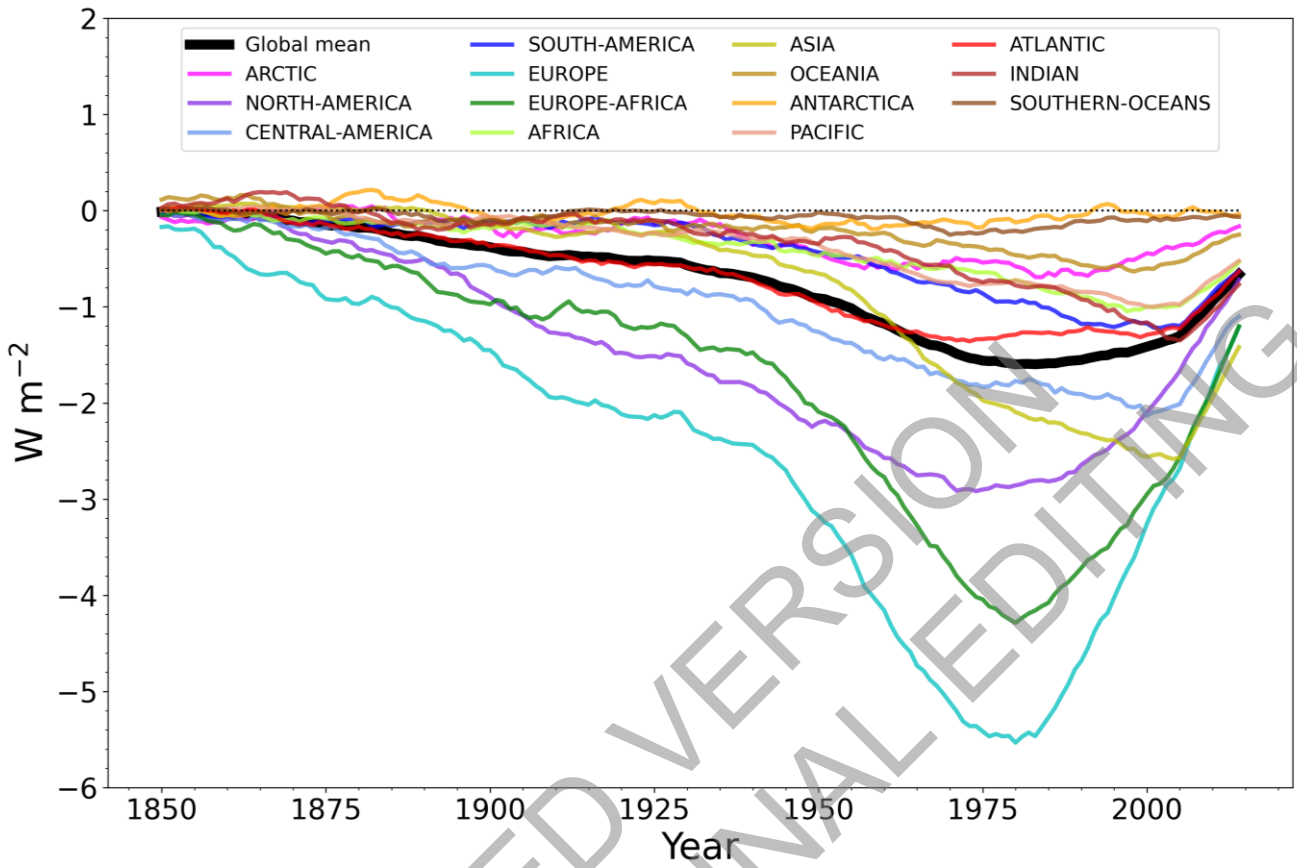
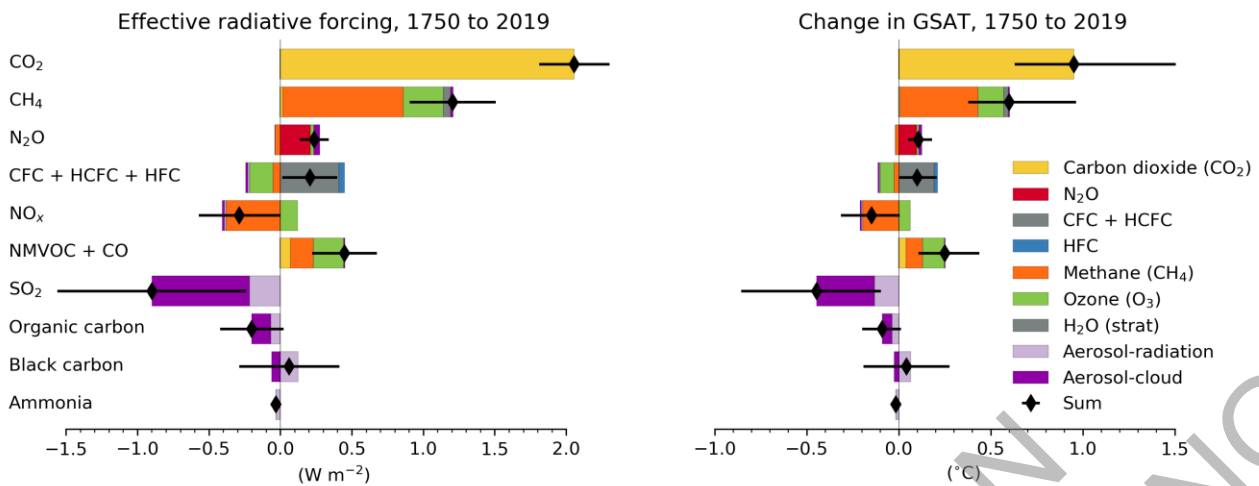


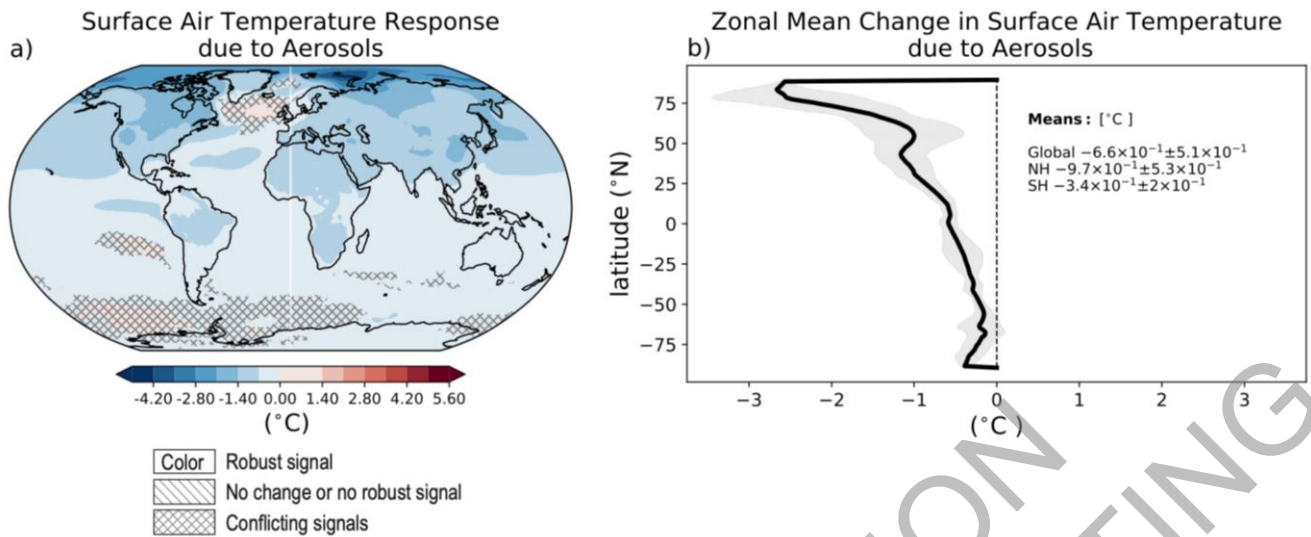
Figure 6.11: Time evolution of 20-year multi-model mean averages of the annual area-weighted mean regional net Effective Radiative Forcings (ERFs) due to aerosols for each of the 14 major regions in the Atlas, and global mean, using the models and model experiments as in Figure 6.10. Further details on data sources and processing are available in the chapter data table (Table 6.SM.1).

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4 **Figure 6.12: Contribution to effective radiative forcing (ERF) (left) and global mean surface air temperature**
5 **(GSAT) change (right) from component emissions between 1750 to 2019** based on CMIP6 models
6 (Thornhill et al., 2021). ERFs for the direct effect of well-mixed greenhouse gases (WMGHGs) are from
7 the analytical formulae in section 7.3.2, H₂O (strat) is from Table 7.8. ERF for other components are
8 multi-model means from Thornhill et al. (2021) and are based on ESM simulations in which emissions of
9 one species at a time are increased from 1850 to 2014 levels. The derived emission-based ERFs are
10 rescaled to match the concentration-based ERFs in Figure 7.6. Error bars are 5-95% and for the ERF
11 account for uncertainty in radiative efficiencies and multi-model error in the means. ERF due to aerosol-
12 radiation (ERF_{ari}) and cloud effects are calculated from separate radiation calls for clear-sky and aerosol
13 free conditions (Ghan, 2013; Thornhill et al., 2021). “Cloud” includes cloud adjustments (semi-direct
14 effect) and ERF from indirect aerosol-cloud to -0.22 W m⁻² for ERF_{ari} and -0.84 W m⁻² interactions
15 (ERF_{aci}). The aerosol components (SO₂, organic carbon, black carbon) are scaled to sum to -0.22 W m⁻²
16 for ERF_{ari} and -0.84 W m⁻² for “cloud” (section 7.3.3). For GSAT estimates, time series (1750-2019) for
17 the ERFs have been estimated by scaling with concentrations for WMGHGs and with historical emissions
18 for SLCFs. The time variation of ERF_{aci} for aerosols is from Chapter 7. The global mean temperature
19 response is calculated from the ERF time series using an impulse response function (see Cross-Chapter
20 Box 7.1) with a climate feedback parameter of -1.31 W m⁻² °C⁻¹. Further details on data sources and
21 processing are available in the chapter data table (Table 6.SM.1).
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3 **Figure 6.13: Multi-model mean surface air temperature response due to aerosol changes between 1850 and**
4 **recent-past (1995-2014)** calculated as the difference between CMIP6 ‘historical’ and AerChemMIP
5 ‘hist-piAer’ experiments, where a) is the spatial pattern of the annual mean surface air temperature
6 response, and b) is the mean zonally averaged response. Model means are derived from years 1995-2014.
7 Uncertainty is represented using the advanced approach. No overlay indicates regions with robust signal,
8 where $\geq 66\%$ of models show change greater than variability threshold and $\geq 80\%$ of all models agree on
9 sign of change; diagonal lines indicate regions with no change or no robust signal, where $< 66\%$ of models
10 show a change greater than the variability threshold; crossed lines indicate regions with conflicting signal,
11 where $\geq 66\%$ of models show change greater than variability threshold and $< 80\%$ of all models agree on
12 sign of change. For more information on the advanced approach, please refer to the Cross-Chapter Box
13 Atlas.1. AerChemMIP models MIROC6, MRI-ESM2-0, NorESM2-LM, GFDL-ESM4, GISS-E2-1-G,
14 UKESM1-0-LL are used in the analysis. Further details on data sources and processing are available in
15 the chapter data table (Table 6.SM.1).
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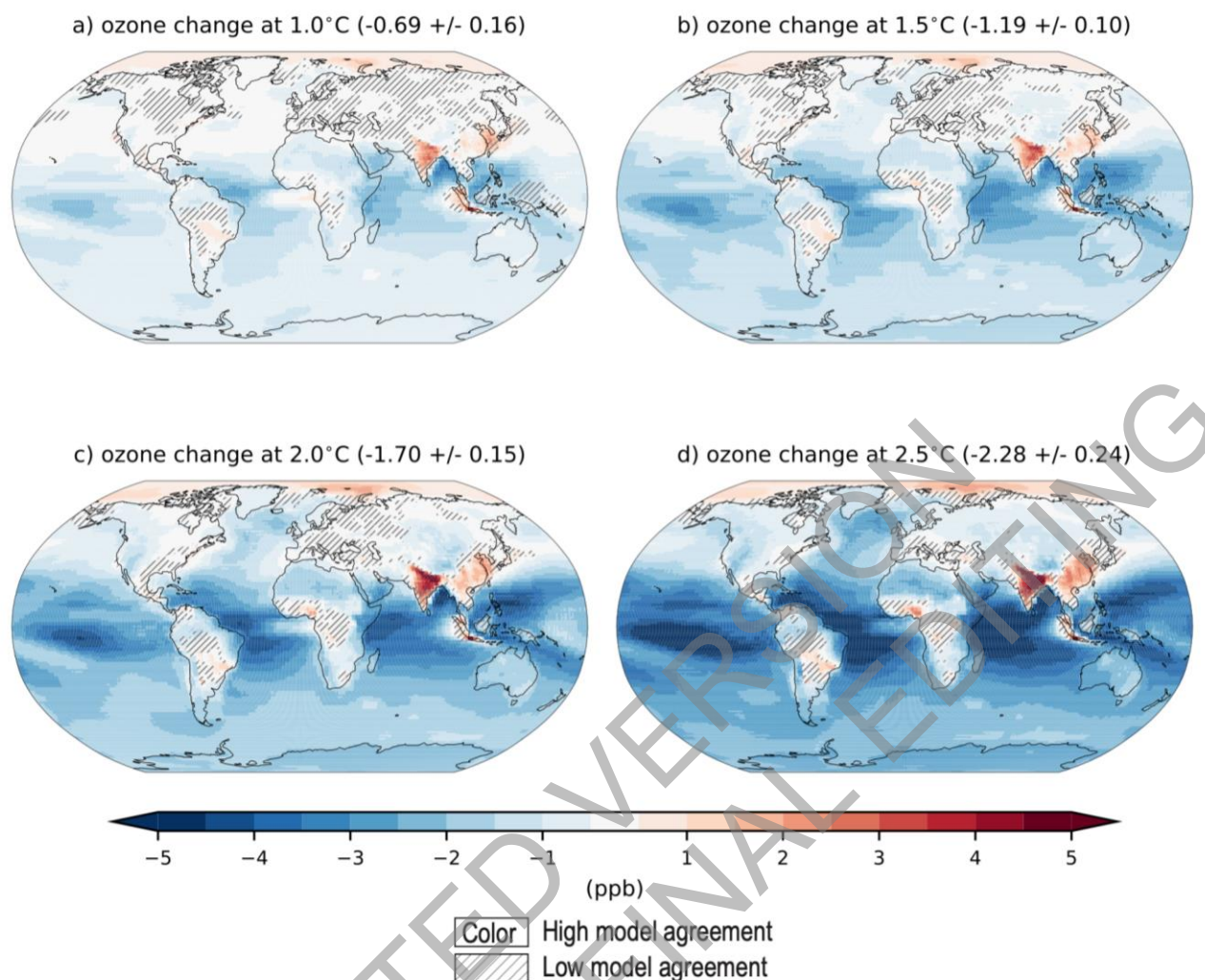
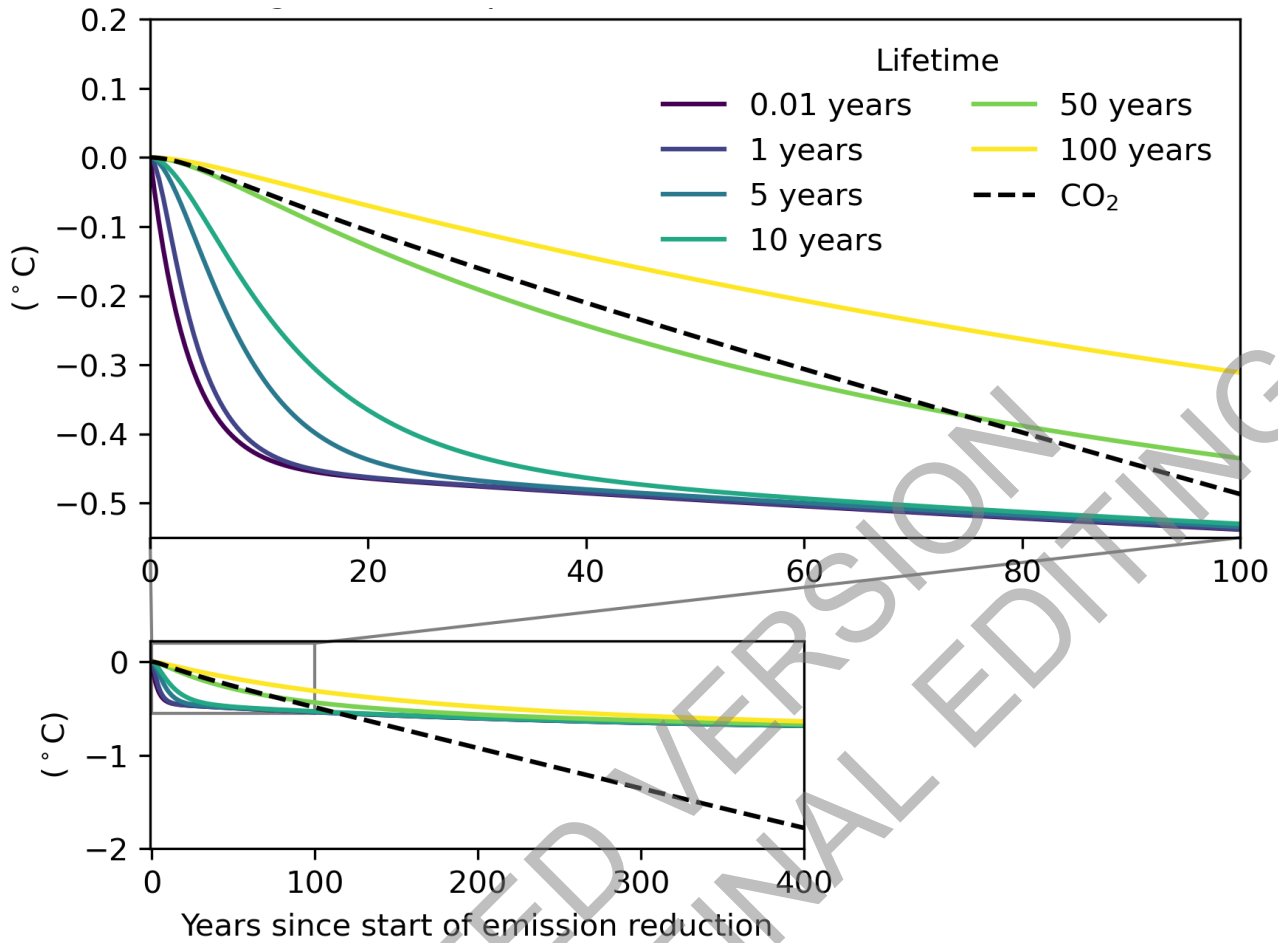


Figure 6.14: Multi-model annual mean change in surface O₃ (ppb) concentrations at different warming levels.

Changes are shown for a) 1.0°C, b) 1.5°C, c) 2.0°C and d) 2.5°C increase in global mean surface air temperature. CMIP6 models include GFDL-ESM4, GISS-E2-1-G, MRI-ESM2-0 and UKESM1-0-LL. For each model the change in surface O₃ is calculated as difference between the ssp370SST and ssp370pdSST experiments in the year when the difference in the global mean surface air temperature between the experiments exceeds the temperature threshold. The difference is calculated as a 20-year mean in surface O₃ around the year when the temperature threshold in each model is exceeded. The multi-model change in global annual mean surface O₃ concentrations with $\pm 1 \sigma$ are shown within parenthesis. Uncertainty is represented using the simple approach: No overlay indicates regions with high model agreement, 3 out of 4 models agree on sign of change; diagonal lines indicate regions with low model agreement, where 3 out of 4 models agree on sign of change. For more information on the simple approach, please refer to the Cross-Chapter Box Atlas.1. Further details on data sources and processing are available in the chapter data table (Table 6.SM.1).

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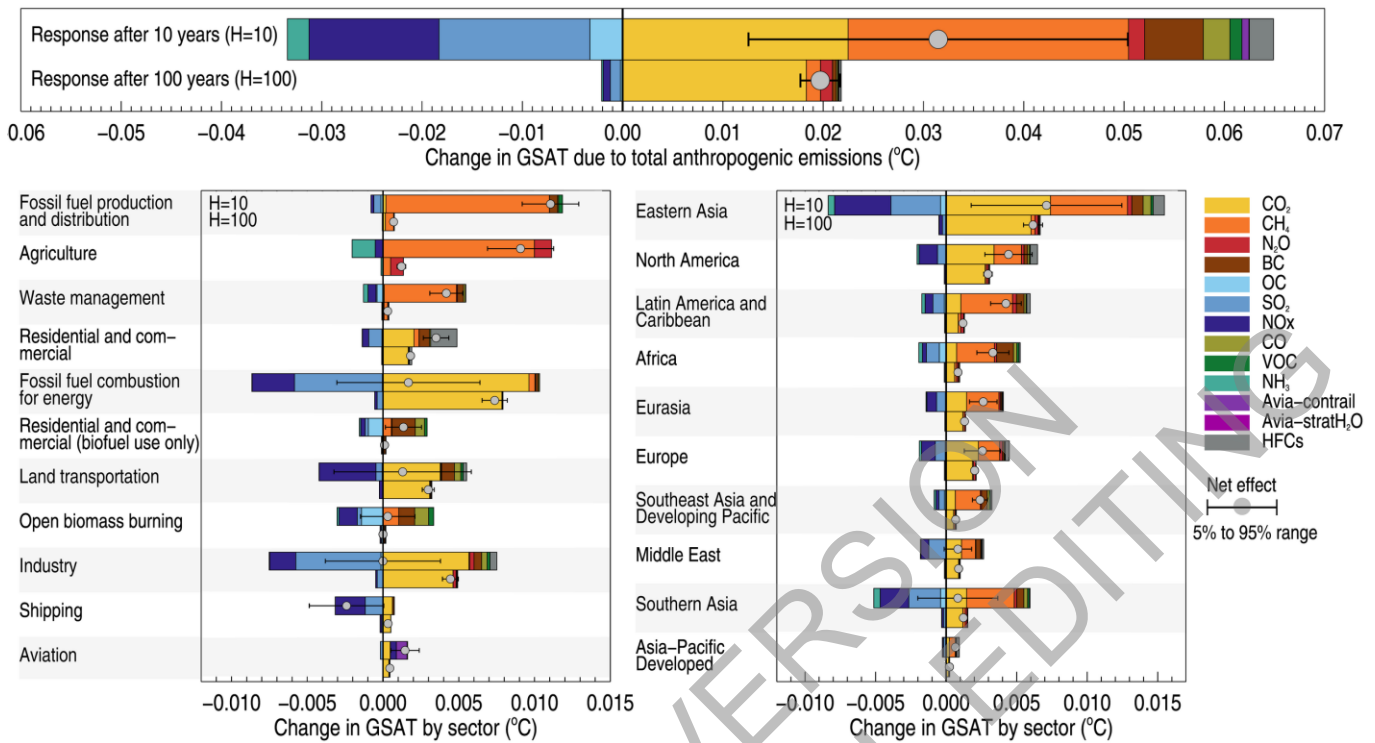


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Figure 6.15: Global mean surface air temperature (GSAT) response to an abrupt reduction in emissions (at time $t=0$) of idealized climate forcing agents with different lifetimes. All emissions are cut to give a radiative forcing of -1 W m^{-2} at steady state (except for CO_2). In other words, if the yearly emissions are E_0 before the reduction, they will have a fixed lower value $E_{\text{year}>0} = (E_0 - \Delta E)$ for all succeeding years. For comparison, the GSAT response to a sustained reduction in CO_2 emissions resulting in an RF of -1 W m^{-2} in year 100 is included (dashed line). The temperature response is calculated using an impulse response function (see Cross-Chapter Box 7.1) with a climate feedback parameter of $-1.31 \text{ W m}^{-2} \text{ }^\circ\text{C}^{-1}$. Further details on data sources and processing are available in the chapter data table (Table 6.SM.1).

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Effect of a one year pulse of present-day emissions on global surface temperature

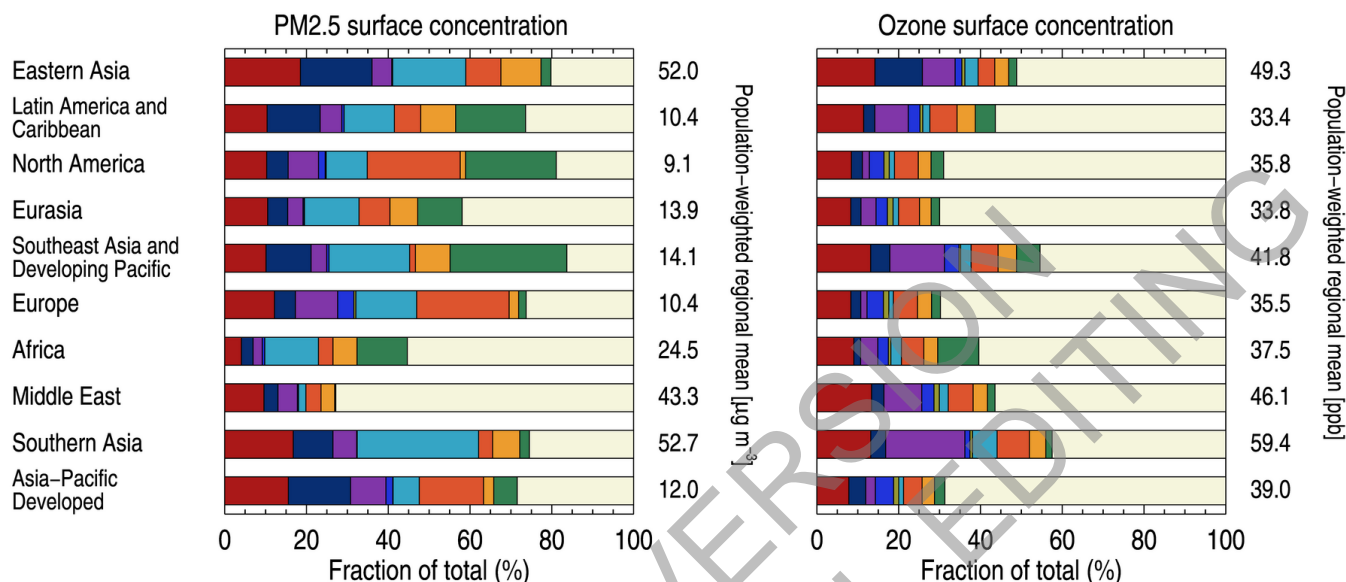
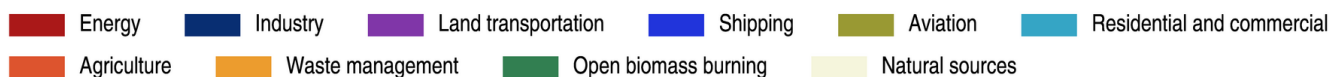


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Figure 6.16: Global-mean temperature response 10 and 100 years following one year of present-day (year 2014) emissions. The temperature response is broken down by individual species and shown for total anthropogenic emissions (top), sectoral emissions (left) and regional emissions (right). Sectors and regions are sorted by (high-to-low) net temperature effect on the 10-year time scale. Error bars in the top panel show uncertainty (5-95% interval) in net temperature effect due to uncertainty in radiative forcing *only* (calculated using a Monte Carlo approach and best estimate uncertainties from the literature - see Lund et al. (Lund et al., 2020) for details). CO₂ emissions are excluded from open biomass burning and residential biofuel use due to their unavailability in CEDS and uncertainties around non-sustainable emission fraction. Emissions for 2014 originate from the Community Emissions Data System (CEDS) (Hoesly et al., 2018), except for HFCs which are from Purohit et al. (2020), open biomass burning from van Marle et al. (2017), and aviation H₂O which is from Lee et al.(2020). The split of fossil fuel production and distribution and combustion for energy and residential and commercial fuel use into fossil fuel and biofuel components obtained from the GAINS model (ECLIPSE version 6b dataset). Open biomass burning emissions are not included for the regions. Emission are aggregated into fossil fuel production and distribution (coal mining, oil and gas production, upstream gas flaring, gas distribution networks), agriculture (livestock and crop production), fossil fuel combustion for energy (power plants), industry (combustion and production processes, solvent use losses from production and end use), residential and commercial (fossil fuel use for cooking and heating as well is HFCs leakage from A/C and refrigeration), waste management (solid waste, incl. landfills and open trash burning, residential and industrial waste water), transport (road and off-road vehicles, and HFC leakage from A/C and refrigeration equipment), residential and commercial (biofuels use for cooking and heating), open biomass burning (forest, grassland, savannah fires, and agricultural waste burning), shipping (incl. international shipping), and aviation (incl. international aviation). Further details on data sources and processing are available in the chapter data table (Table 6.SM.1).

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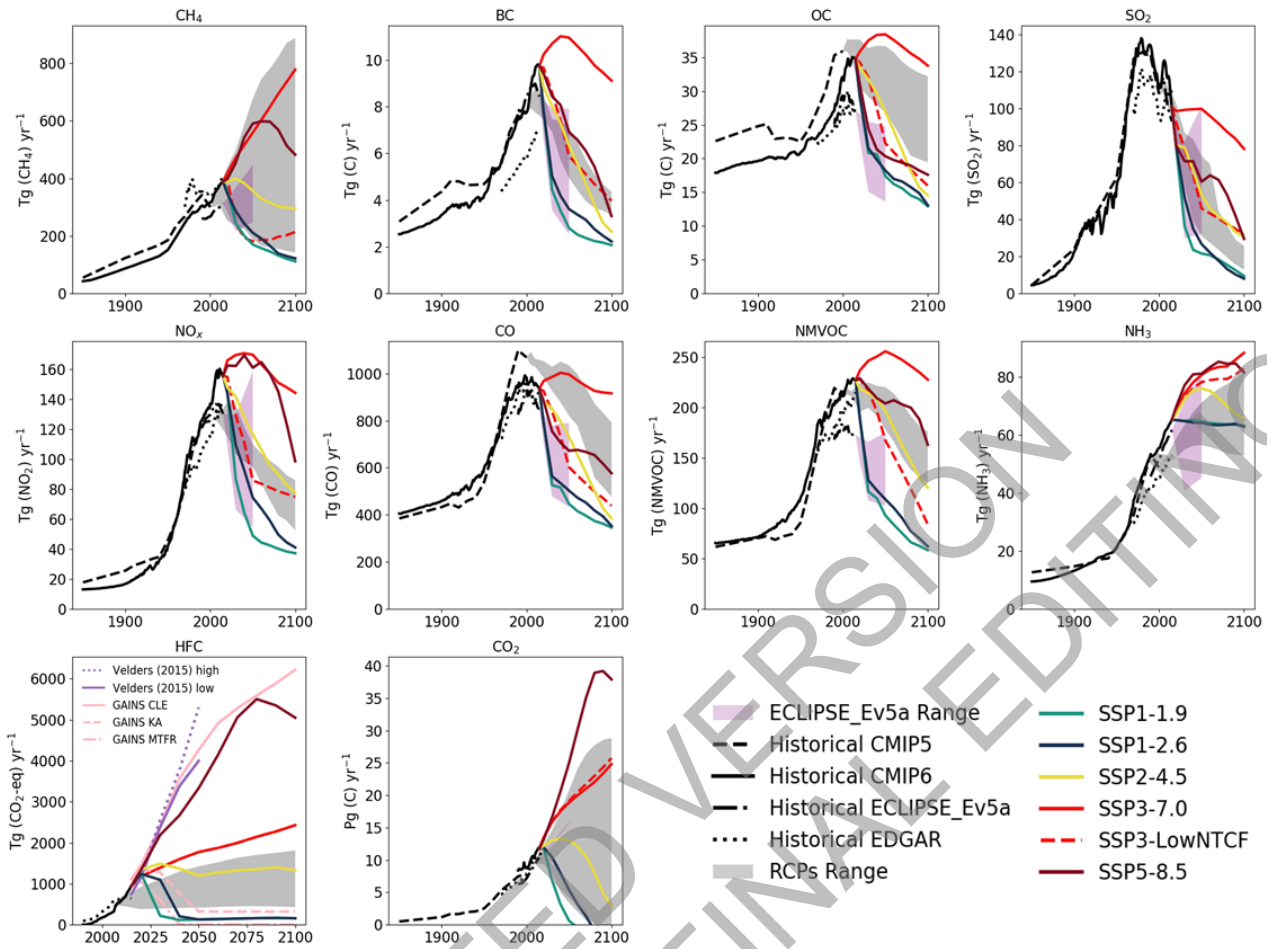
Attribution of regional, population-weighted PM2.5 and ozone to sectors



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Figure 6.17: Emission source-sector attribution of regional population weighted mean concentrations of PM_{2.5} and ozone for present day emissions (year 2014). Regional concentrations and source apportionment calculated with the TM5-FASST model (Van Dingenen et al., 2018) for the 2014 emission data from the Community Emissions Data System (CEDS) (Hoesly et al., 2018) and van Marle et al.(2017) for open biomass burning. PM_{2.5} dust and seasalt are monthly mean climatological average over 2010-2018 from CAMS global reanalysis (EAC4) (Inness et al., 2019), generated using Copernicus Climate Change Service information [January 2020]. Anthropogenic sectors are similar to those in Figures 6.2 and 6.16 except grouping of fossil fuel production, distribution and combustion for energy under “Energy” and grouping of use of fossil fuel and biofuel use for cooking and heating under “Residential and Commercial”. Further details on data sources and processing are available in the chapter data table (Table 6.SM.1).

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Figure 6.18: Global anthropogenic and biomass burning short-lived climate forcer (SLCF) and CO₂ emissions from 1850 to 2100 and HFC emissions from 1990 to 2100.

Emissions for the Coupled Model Intercomparison Project Phase 6 (CMIP6) for the period 1850-2014 are based on Hoesly et al. (2018) and van Marle et al. (2017), emissions for CMIP5 for the period 1850-2005 are from Lamarque et al. (2010), CO₂ emissions are from EDGAR database (Crippa et al., 2020); CH₄ and HFCs are from (Crippa et al., 2019); and air pollutants are from (EC-JRC / PBL, 2020), Höglund-Isaksson (2012) and Klimont et al. (2017) for ECLIPSE. Projections originate from the Shared Socio-Economic Pathway (SSP) database (Riahi et al., 2017; Rogelj et al., 2018; Gidden et al., 2019), Representative Concentration Pathway (RCP) database (van Vuuren et al., 2011), GAINS (CLE – current legislation baseline without Kigali Amendment, KA – Kigali Amendment, MTR – maximum technical mitigation potential) for HFCs (Purohit et al., 2020), Velders et al. (2015), ECLIPSE (Stohl et al., 2015). Further details on data sources and processing are available in the chapter data table (Table 6.SM.1).

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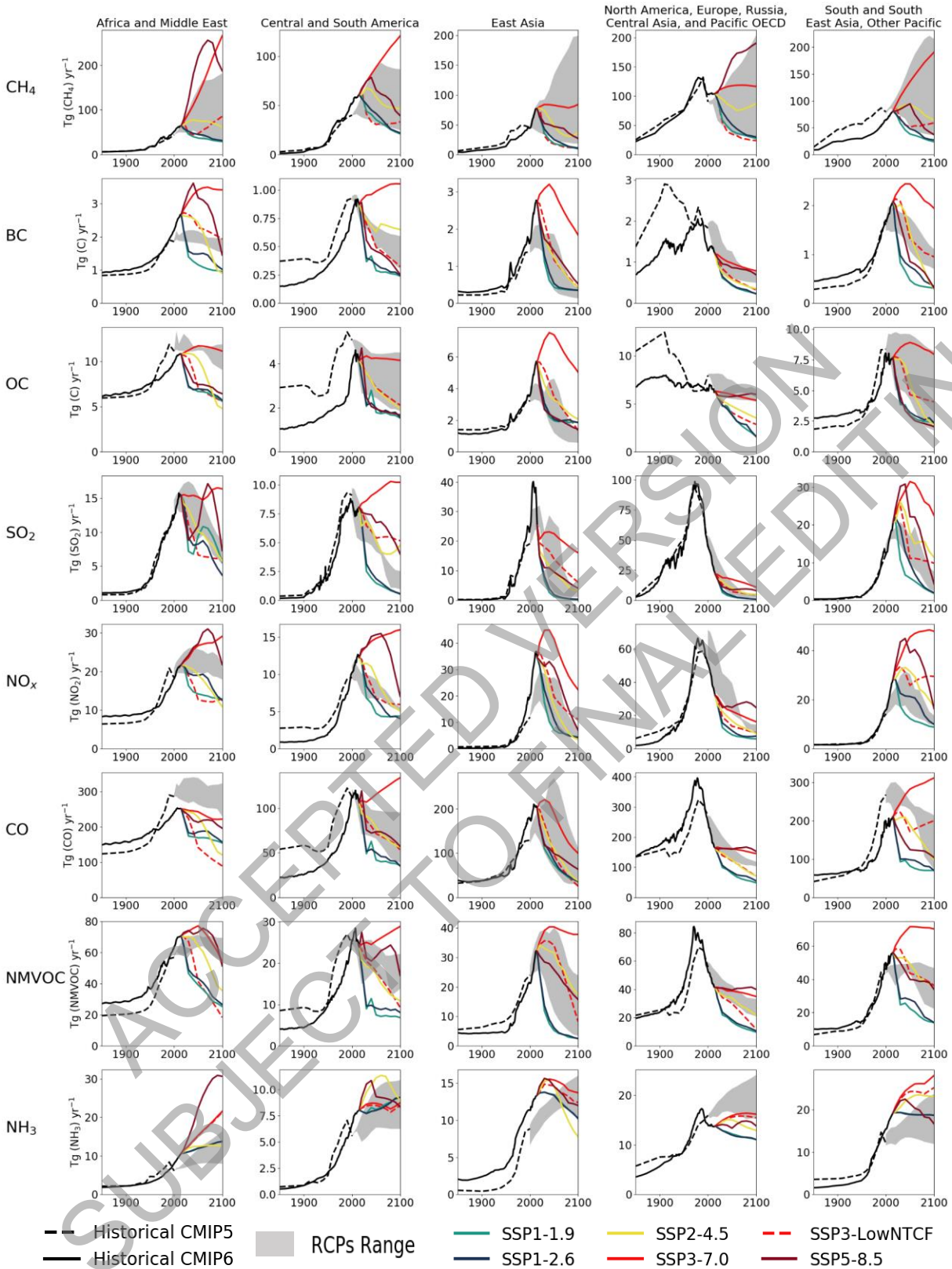
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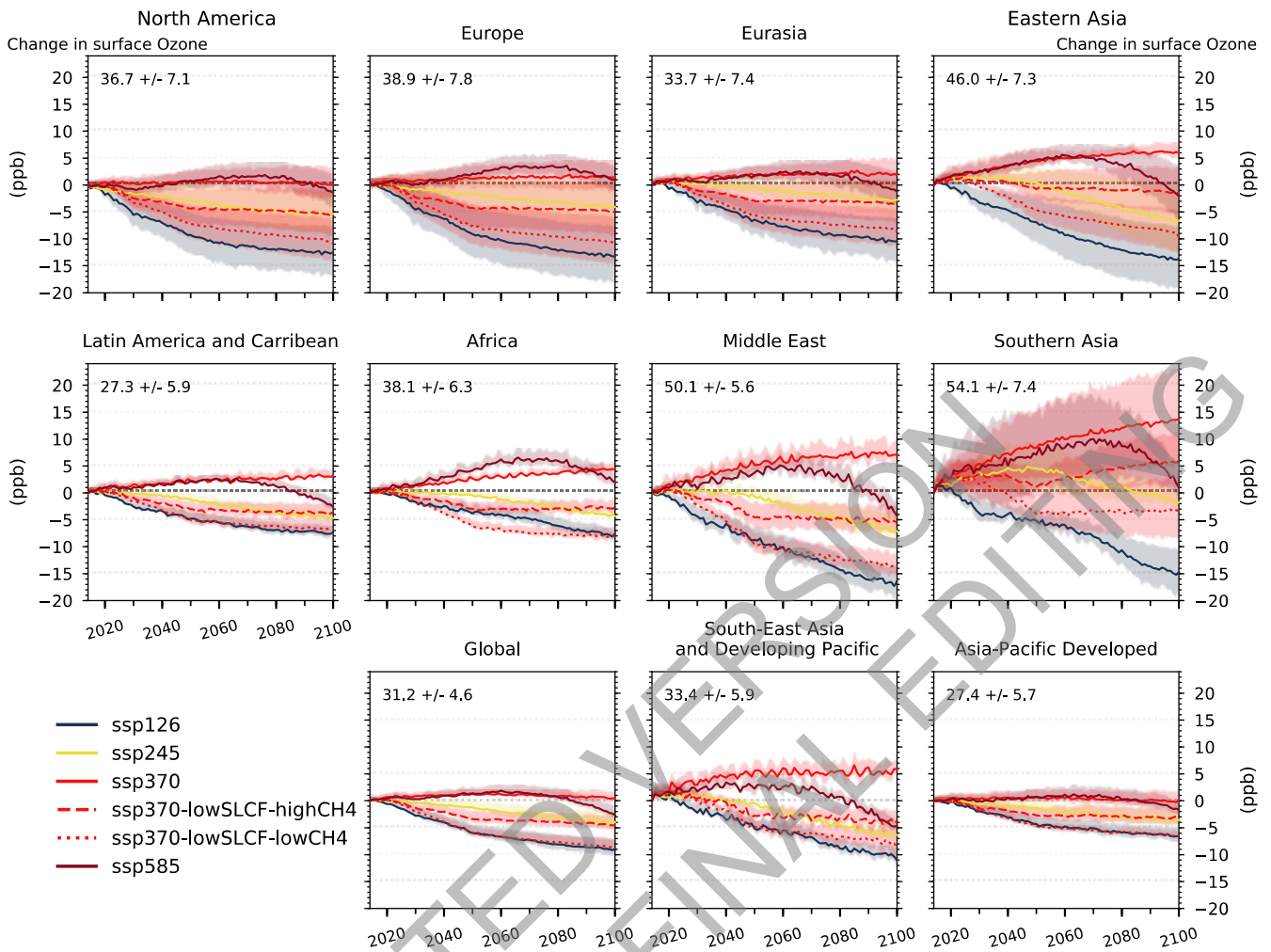
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Figure 6.19: Regional anthropogenic and biomass burning short-lived climate forcer (SLCF) emissions from 1850 to 2100. Emissions for the Coupled Model Intercomparison Project Phase 6 (CMIP6) for the period 1850-2014 are based on Hoesly et al. (2018) and van Marle et al. (2017) and emissions for CMIP5 for the period 1850-2005 are from Lamarque et al. (2010). Projections originate from the Shared Socio-

1 Economic Pathway (SSP) database (Riahi et al., 2017; Rogelj et al., 2018; Gidden et al., 2019) and
2 Representative Concentration Pathway (RCP) database (van Vuuren et al., 2011). Further details on data
3 sources and processing are available in the chapter data table (Table 6.SM.1).
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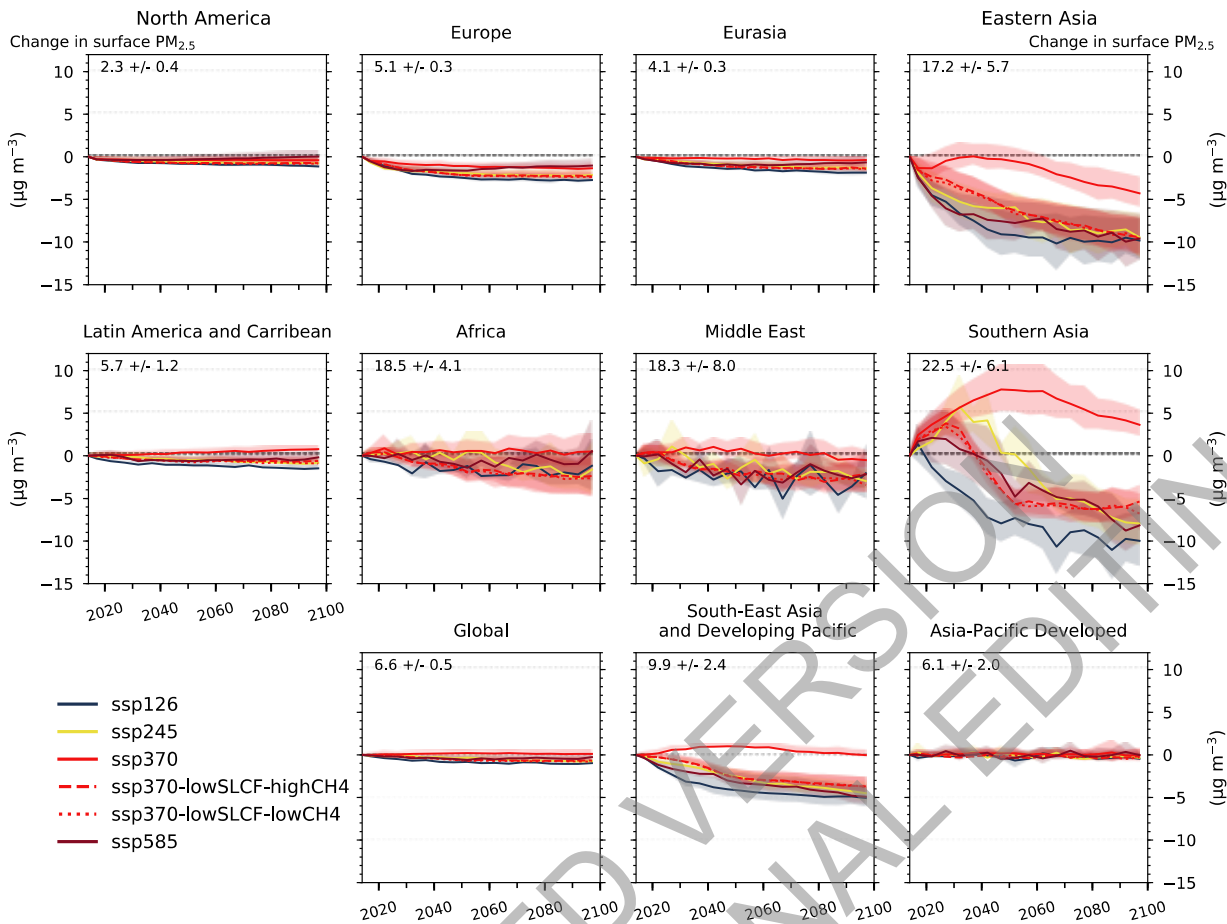
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Figure 6.20: Projected changes in regional annual mean surface ozone (O₃) (ppb) from 2015 to 2100 in different shared socio-economic pathways (SSPs). Each panel represents values averaged over the corresponding land area (except for Global) shown on the map in Figure 6.7. Solid colored lines and shading indicate the multi-model mean and $\pm 1 \sigma$ across the available CMIP6 models (Turnock et al. 2020; Allen et al. 2021) for each scenario. Changes are relative to annual mean values calculated over the period 2005-2014 from the historical experiment as indicated in the top left of each regional panel along with $\pm 1 \sigma$. For each model all available ensemble members are averaged before being used to calculate the multi-model mean. Ozone changes are also displayed in the Interactive Atlas. Further details on data sources and processing are available in the chapter data table (Table 6.SM.1).

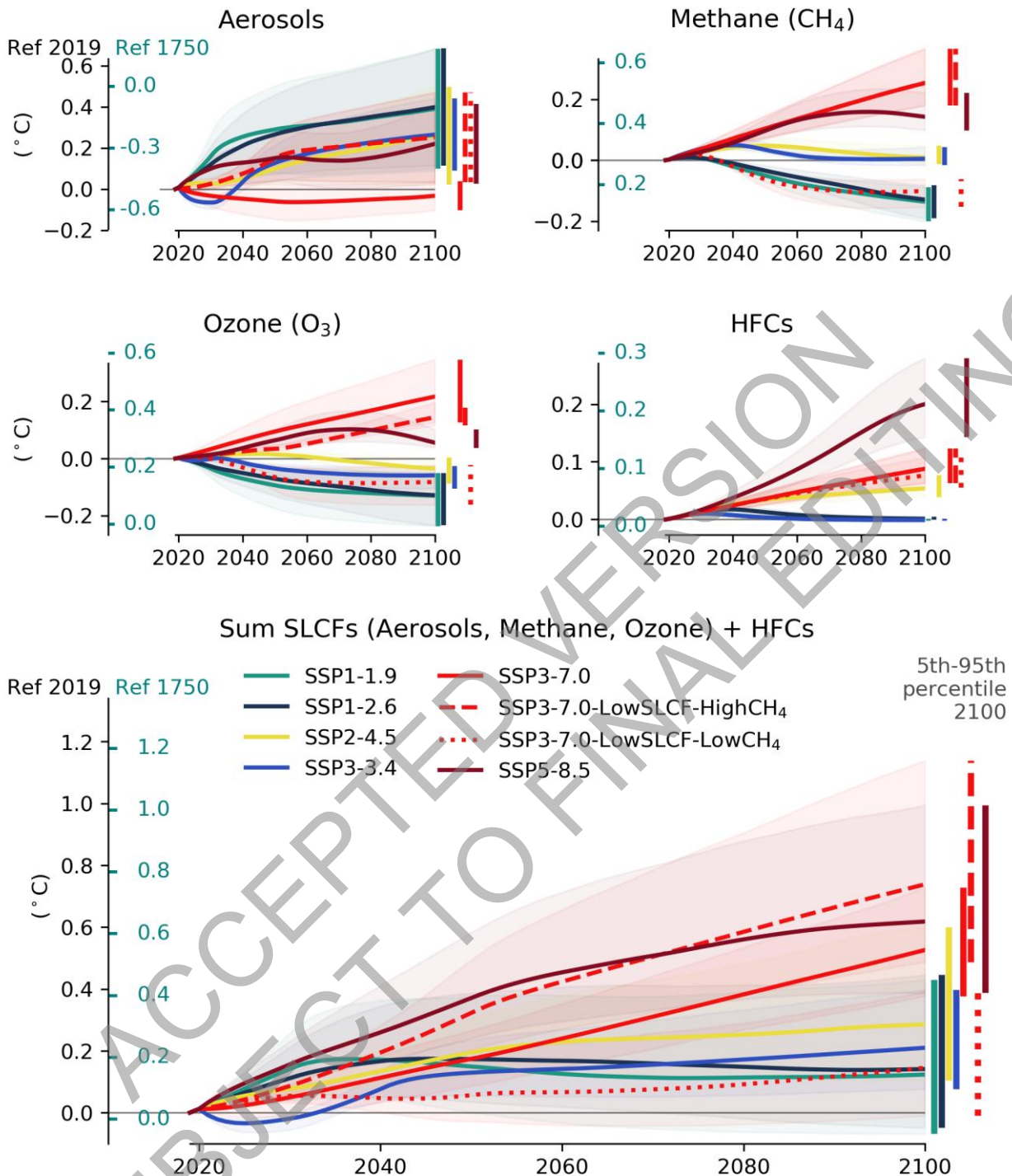
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Figure 6.21: Future changes in regional 5-year mean surface PM_{2.5} from 2015 to 2100 in different shared socio-economic pathways (SSPs). PM_{2.5} stands for micrograms per cubic meter of aerosols with diameter less than 2.5 µm and is calculated by summing up individual aerosol mass components from each model as: black carbon + particulate organic matter + sulphate + 0.25 * sea salt + 0.1 * dust. Since not all CMIP6 models reported nitrate aerosol, it is not included here. See Figure 6.20 for further details. PM_{2.5} changes are also displayed in the Interactive Atlas. Further details on data sources and processing are available in the chapter data table (Table 6.SM.1).

Effect on Global Surface Air Temperature (GSAT) relative to 2019



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Figure 6.22: Time evolution of the effects of short-lived climate forcers (SLCFs) and hydrofluorocarbons (HFCs) on global surface air temperature (GSAT) across the WGI core set of Shared Socio-Economic Pathways (SSPs). Effects of net aerosols, methane, tropospheric ozone and hydrofluorocarbons (HFCs) (with lifetimes <50years), and the sum of these, relative to year 2019 and to year 1750. The GSAT changes are based on the assessed historic and future evolution of Effective Radiative Forcing (Section 7.3.5). The temperature responses to the ERFs are calculated with an impulse response function with an equilibrium climate sensitivity of 3.0°C for a doubling of atmospheric CO₂ (feedback parameter of -1.31 W m⁻² °C⁻¹, see CC-Box 7.1). The vertical bars to the right in each panel show the uncertainties (5-95% ranges) for the GSAT change between 2019 and 2100. Further details on data sources and processing are available in the chapter data table (Table 6.SM.1).

Contribution from changes in anthropogenic regional emissions of SLCFs to GSAT changes

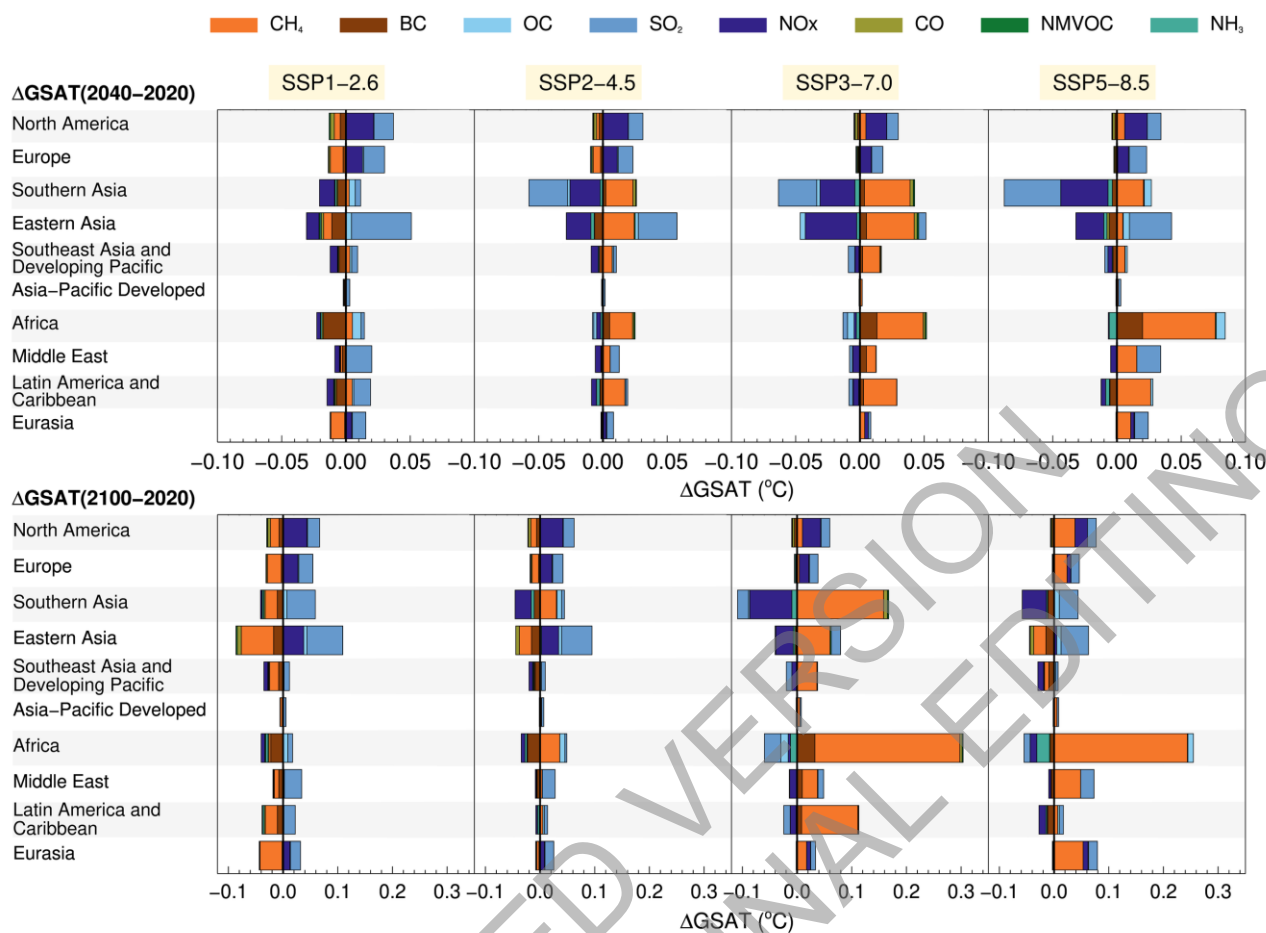


Figure 6.23: Contribution from regional emissions of short-lived climate forcers (SLCFs) to changes in global surface air temperature (GSAT) in 2040 (upper row) and 2100 (lower row), relative to 2020 for four Shared Socio-economic Pathways (SSP). Adapted from (Lund et al., 2020). NO_x , CO, and VOC account for the impact through changes in ozone and methane, NO_x additionally includes the impact through formation of nitrate aerosols. BC, SO_2 and OC accounts for the direct aerosol effect (aerosol-radiation interactions), as well as an estimate of the semi-direct effect for BC due to rapid adjustments and indirect effect (aerosol-cloud interactions) of sulfate aerosols. Regions are the same as shown in the map in Figure 6.7. Further details on data sources and processing are available in the chapter data table (Table 6.SM.1).

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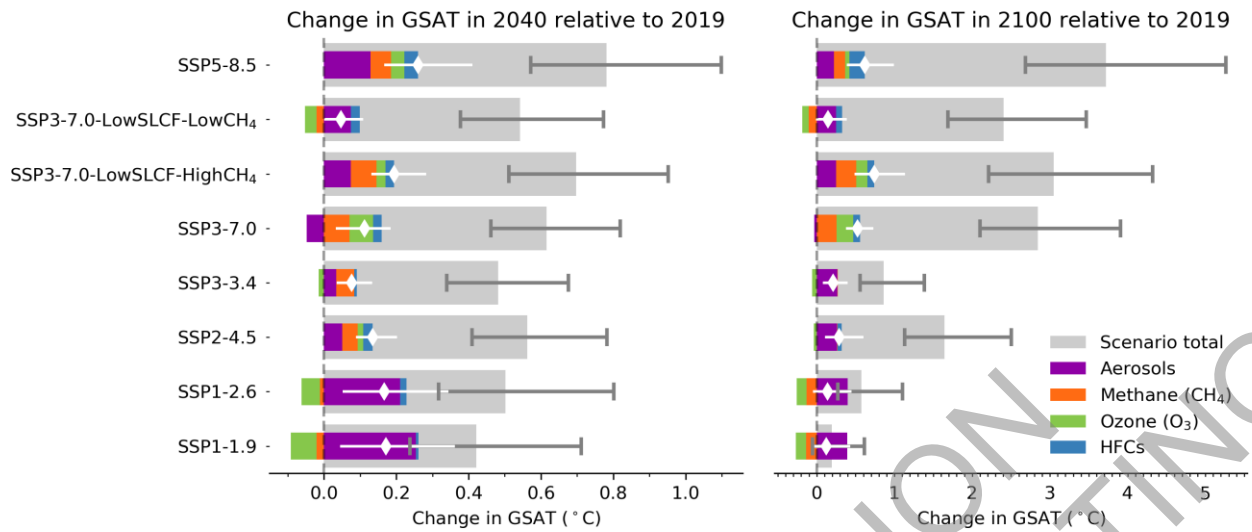
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Figure 6.24: Effects of short-lived climate forcers (SLCFs) and hydrofluorocarbons (HFCs) on global surface air temperature (GSAT) across the WGI core set of Shared Socio-Economic Pathways (SSPs). Effects of net aerosols, methane, tropospheric ozone, and hydrofluorocarbons (HFCs; with lifetimes <50years), are compared with those of total anthropogenic forcing for 2040 and 2100 relative to year 2019. The GSAT changes are based on the assessed historic and future evolution of Effective Radiative Forcing (Section 7.3.5). The temperature responses to the ERFs are calculated with an impulse response function with an equilibrium climate sensitivity of 3.0°C for a doubling of atmospheric CO_2 (feedback parameter of $-1.31 \text{ W m}^{-2} \text{ }^{\circ}\text{C}^{-1}$, see Cross-Chapter Box 7.1). Uncertainties are 5-95% ranges. The scenario total (grey bar) includes all anthropogenic forcings (long- and short-lived climate forcers, and land use changes) whereas the white points and bars show the net effects of SLCFs and HFCs and their uncertainties. Further details on data sources and processing are available in the chapter data table (Table 6.SM.1).

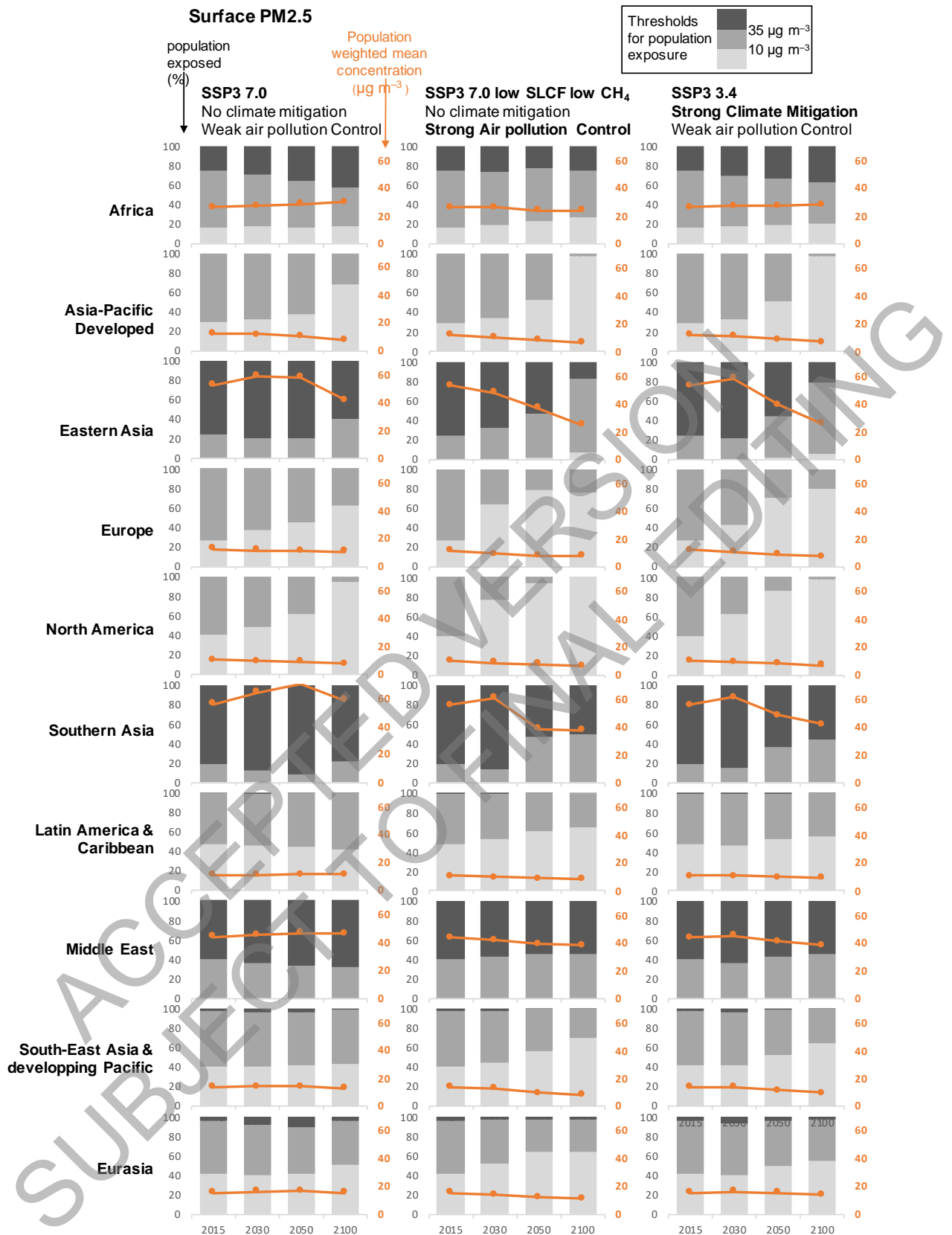


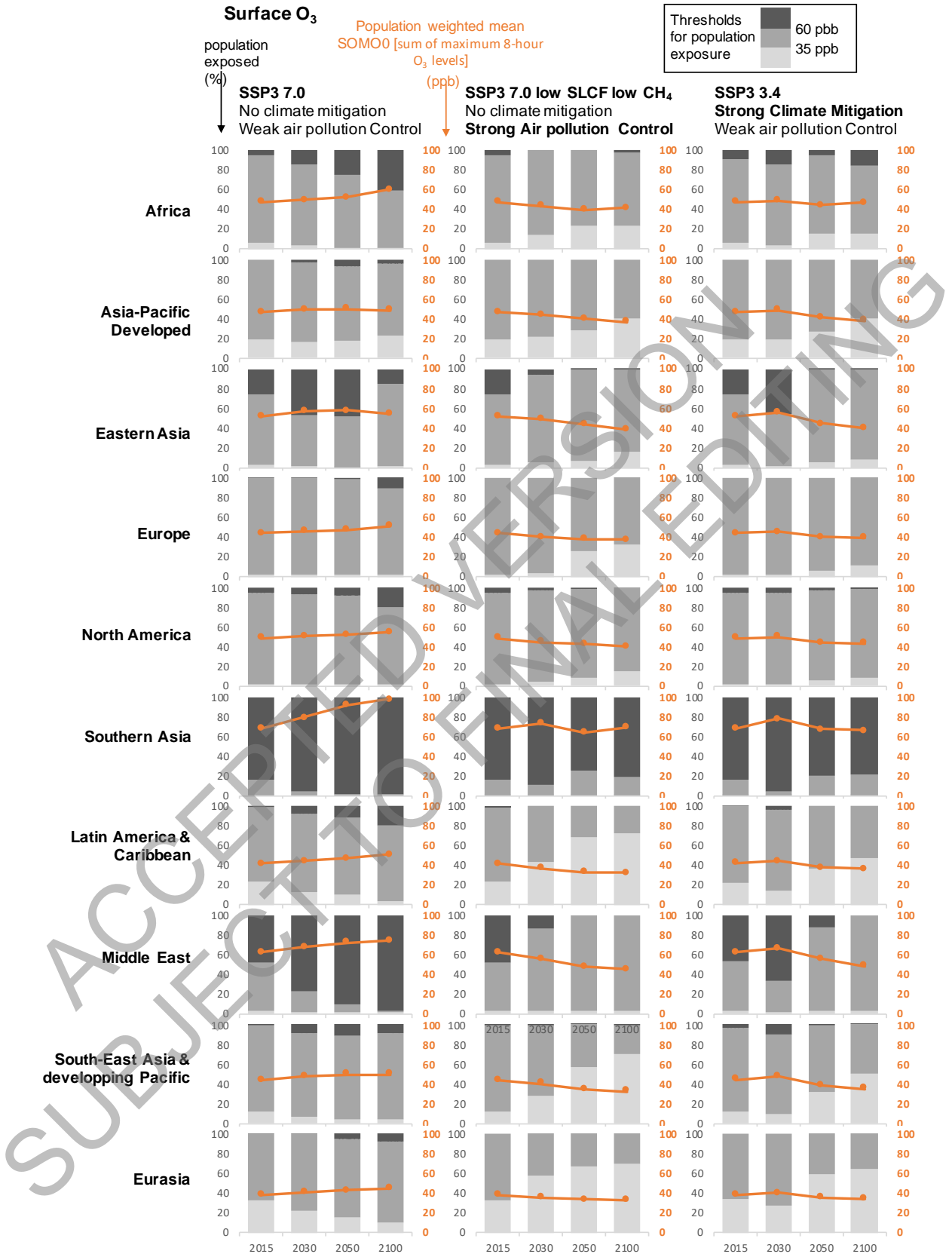
Figure 6.25: Effect of dedicated air pollution or climate policy on population-weighted PM_{2.5} concentrations ($\mu\text{g m}^{-3}$) and share of population (%) exposed to different PM_{2.5} levels across selected world regions. Thresholds of $10 \mu\text{g m}^{-3}$ and $35 \mu\text{g m}^{-3}$ represent the WHO air quality guideline and the WHO interim target 1, respectively; WHO (2017). Results are compared for SSP3-7.0 (no major improvement of current legislation is assumed), SSP3-LowSLCF (strong air pollution controls are assumed), and a climate

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1 mitigation scenario SSP3-3.4; details of scenario assumptions are discussed in Riahi et al. (2017) and Rao
2 et al. (2017). Analysis performed with the TM5-FASST model (Van Dingenen et al., 2018) using
3 emission projections from the Shared Socio-Economic Pathway (SSP) database (Riahi et al., 2017;
4 Rogelj et al., 2018; Gidden et al., 2019). Further details on data sources and processing are available in
5 the chapter data table (Table 6.SM.1).
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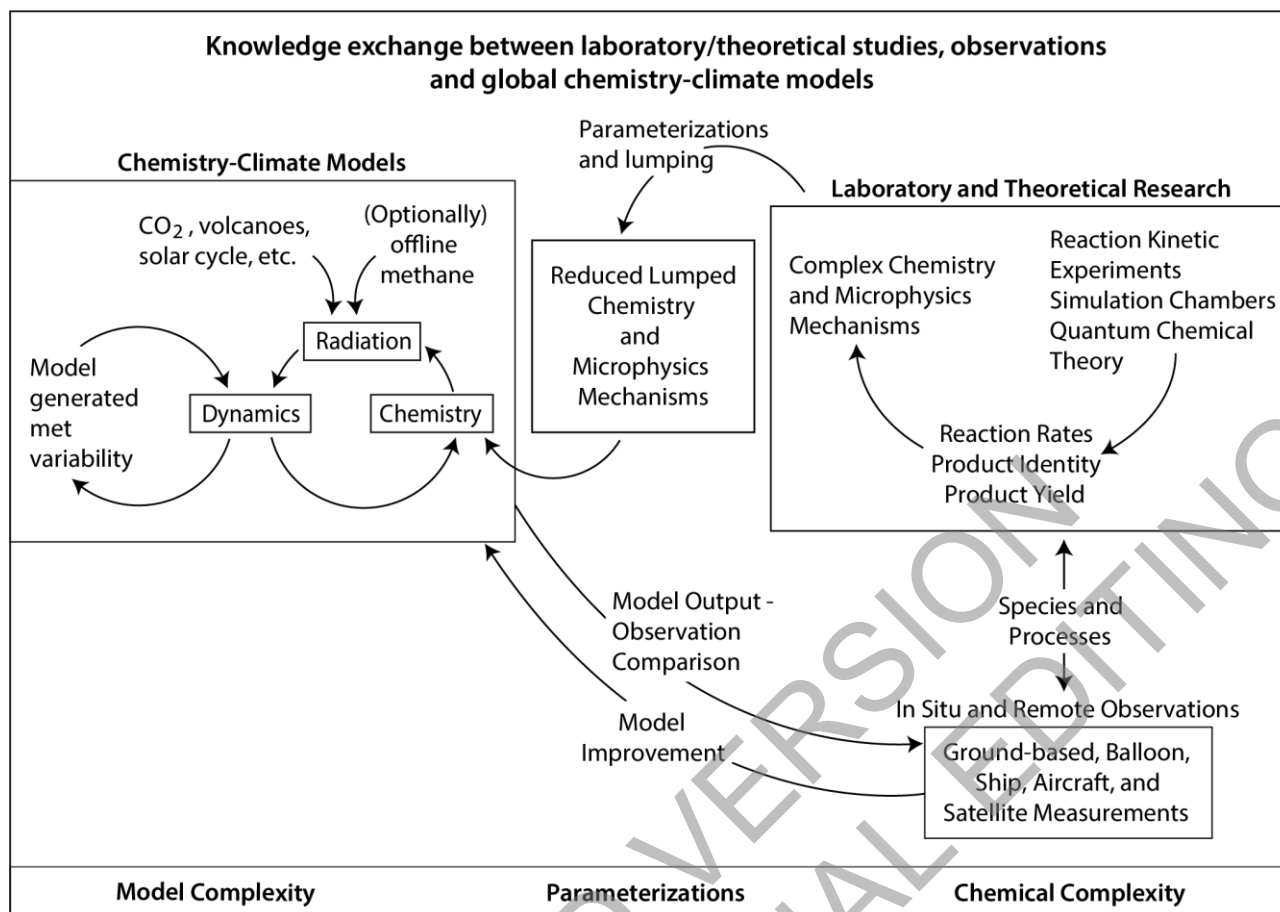
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Figure 6.26: Effect of dedicated air pollution or climate policy on population-weighted ozone concentrations (SOMO0; ppb) and share of population (%) exposed to chosen ozone levels across ten world regions. Results are compared for SSP3-7.0 (no major improvement of current legislation is assumed),

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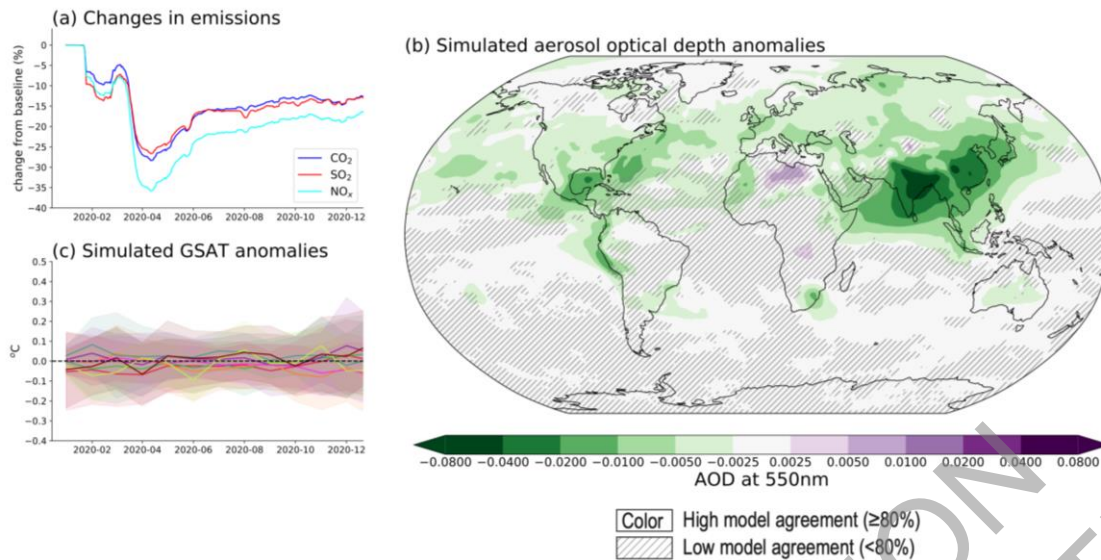
SSP3-LowSLCF (strong air pollution controls are assumed), and a climate mitigation scenario (SSP3-3.4); details of scenario assumptions are discussed in Riahi et al. (2017) and Rao et al. (2017). Analysis performed with the TM5-FASST model (Van Dingenen et al., 2018) using emission projections from the Shared Socio-Economic Pathway (SSP) database [<https://tntcat.iiasa.ac.at/SspDb/dsd>] (Riahi et al., 2017; Rogelj et al., 2018; Gidden et al., 2019). Further details on data sources and processing are available in the chapter data table (Table 6.SM.1).

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Box 6.1, Figure 1: Knowledge exchange between laboratory/theoretical studies, observations and global chemistry-climate models to inform our understanding of short-lived climate forcers (SLCFs).

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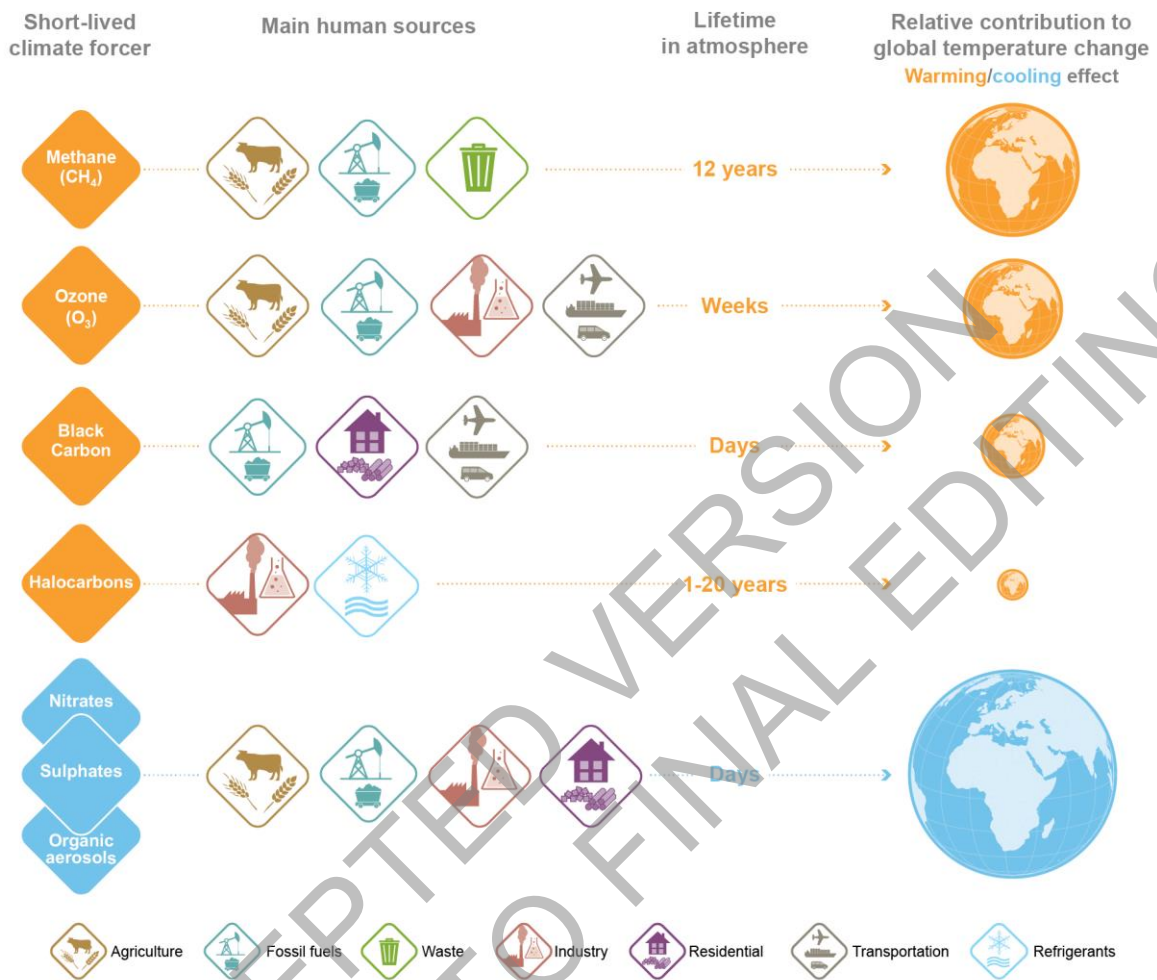
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Cross-Chapter Box 6.1, Figure 1: Emission reductions and their effect on aerosols and climate in response to COVID-19. Estimated reductions in emissions of CO₂, SO₂ and NO_x are shown in panel (a) based on reconstructions using activity data (updated from Forster et al., 2020). Eight Earth System Models (ESMs) performed multiple ensemble simulations of the response to COVID-19 emissions reductions forced with these assumed emission reductions up till August 2020 followed by a constant continuation near the August value to the end of 2020. Emission reductions were applied relative to the SSP2-4.5 scenario. Panel (b) shows ESM simulated AOD at 550nm (only seven models reported this variable). Panel (c) shows ESM simulated GSAT anomalies during 2020; curves denote the ensemble mean result for each model with shading used for ±1 standard deviation for each model. ESM data from these simulations (“ssp245-covid”) is archived on the Earth System Grid CMIP6 database. Uncertainty is represented using the simple approach: No overlay indicates regions with high model agreement, where ≥80% of models agree on sign of change; diagonal lines indicate regions with low model agreement, where <80% of models agree on sign of change. For more information on the simple approach, please refer to the Cross-Chapter Box Atlas.1.

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FAQ 6.1: What are short-lived climate forcers and how do they affect the climate?

Short lived climate forcers do not remain for very long in the atmosphere, thus an increase or decrease in their emissions rapidly affects the climate system.

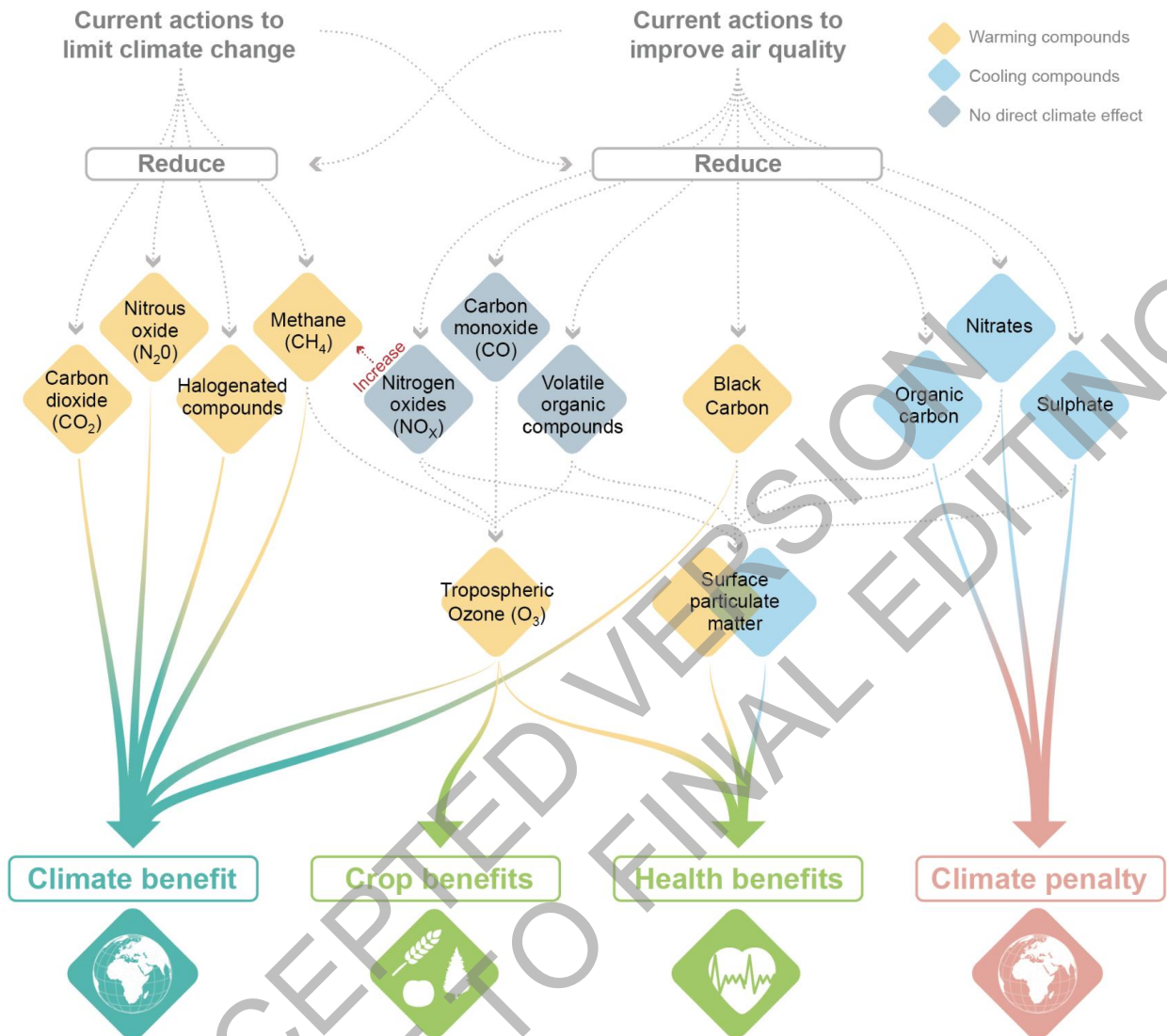


FAQ 6.1, Figure 1: Main short-lived climate forcers, their sources, how long they exist in the atmosphere, and their relative contribution to global surface temperature changes between 1750 and 2019 (area of the globe). By definition, this contribution depends on the lifetime, the warming/cooling potential (radiative efficiency), and the emissions of each compound in the atmosphere. Blue indicates cooling and orange warming. Note that, between 1750 and 2019, the cooling contribution from aerosols (blue diamonds and globe) was approximately half the warming contribution from carbon dioxide.

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FAQ 6.2: Limiting climate change and improving air quality?

Climate change and air quality are so intimately linked that addressing one issue can affect the other one.



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3 **FAQ6.2, Figure 1: Links between actions aiming to limit climate change and actions to improve air**
4 **quality.** Greenhouse gases and aerosols (orange and blue) can affect directly climate. Air
5 pollutants (bottom) can affect the human health, ecosystems and climate. All these compounds
6 have common sources and sometimes interact with each other in the atmosphere which makes
7 impossible to consider them separately (dotted grey arrows).
8