TS.3 Drivers of Climate Change

TS.3.1 Introduction

Human activities have changed and continue to change the Earth's surface and atmospheric composition. Some of these changes have a direct or indirect impact on the energy balance of the Earth and are thus drivers of climate change. Radiative forcing (RF) is a measure of the net change in the energy balance of the Earth system in response to some external perturbation (see Box TS.2), with positive RF leading to a warming and negative RF to a cooling. The RF concept is valuable for comparing the influence on GMST of most individual agents affecting the Earth's radiation balance. The quantitative values provided in AR5 are consistent with those in previous IPCC reports, though there have been some important revisions (Figure TS.6). Effective radiative forcing (ERF) is now used to quantify the impact of some forcing agents that involve rapid adjustments of components of the atmosphere and surface that are assumed constant in the RF concept (see Box TS.2). RF and ERF are estimated from the change between 1750 and 2011, referred to as 'Industrial Era', if other time periods are not explicitly stated. Uncertainties are given associated with the best estimates of RF and ERF, with values representing the 5 to 95% (90%) confidence range. {8.1, 7.1}

In addition to the global mean RF or ERF, the spatial distribution and temporal evolution of forcing, as well as climate feedbacks, play a role in determining the eventual impact of various drivers on climate. Land surface changes may also impact the local and regional climate through processes that are not radiative in nature. {8.1, 8.3.5, 8.6}

TS.3.2 Radiative Forcing from Greenhouse Gases

Human activity leads to change in the atmospheric composition either directly (via emissions of gases or particles) or indirectly (via atmospheric chemistry). Anthropogenic emissions have driven the changes in well-mixed greenhouse gas (WMGHG) concentrations during the Industrial Era (see Section TS.2.8 and TFE.7). As historical WMGHG concentrations since the pre-industrial are well known based on direct measurements and ice core records, and WMGHG radiative properties are also well known, the computation of RF due to concentration changes provides tightly constrained values (Figure TS.6). There has not been significant change in our understanding of WMGHG radiative impact, so that the changes in RF estimates relative to AR4 are due essentially to concentration increases. The best estimate for WMGHG ERF is the same as RF, but the uncertainty range is twice as large due to the poorly constrained cloud responses. Owing to high-quality observations, it is certain that increasing atmospheric burdens of most WMGHGs, especially CO_2, resulted in a further increase in their RF from 2005 to 2011. Based on concentration changes, the RF of all WMGHGs in 2011 is 2.83 [2.54 to 3.12] W m^{-2} (*very high confidence*). This is an increase since AR4 of 0.20 [0.18 to 0.22] W m^{-2}, with nearly all of the increase due to the increase in the abundance of CO_2 since 2005. The Industrial Era RF for CO_2 alone is 1.82 [1.63 to 2.01] W m^{-2}. Over the last 15 years, CO_2 has been the dominant contributor to the increase in RF from the WMGHGs, with RF of CO_2 having an average growth rate slightly less than 0.3 W m^{-2} per decade. The uncertainty in the WMGHG RF is due in part to its radiative properties but mostly to the full accounting of atmospheric radiative transfer including clouds. {2.2.1, 5.2, 6.3, 8.3, 8.3.2; Table 6.1}

After a decade of near stability, the recent increase of CH_4 concentration led to an enhanced RF compared to AR4 by 2% to 0.48 [0.43 to 0.53] W m^{-2}. It is *very likely* that the RF from CH_4 is now larger than that of all halocarbons combined. {2.2.1, 8.3.2}

Atmospheric N_2O has increased by 6% since AR4, causing an RF of 0.17 [0.14 to 0.20] W m^{-2}. N_2O concentrations continue to rise while those of dichlorodifluoromethane (CF_2Cl_2, CFC-12), the third largest WMGHG contributor to RF for several decades, are decreasing due to phase-out of emissions of this chemical under the Montreal Protocol. Since

Box TS.2 | Radiative Forcing and Effective Radiative Forcing

RF and ERF are used to quantify the change in the Earth's energy balance that occurs as a result of an externally imposed change. They are expressed in watts per square metre (W m^{-2}). RF is defined in AR5, as in previous IPCC assessments, as the change in net downward flux (shortwave + longwave) at the tropopause after allowing for stratospheric temperatures to readjust to radiative equilibrium, while holding other state variables such as tropospheric temperatures, water vapour and cloud cover fixed at the unperturbed values (see Glossary). {8.1.1}

Although the RF concept has proved very valuable, improved understanding has shown that including rapid adjustments of the Earth's surface and troposphere can provide a better metric for quantifying the climate response. These rapid adjustments occur over a variety of time scales, but are relatively distinct from responses to GMST change. Aerosols in particular impact the atmosphere temperature profile and cloud properties on a time scale much shorter than adjustments of the ocean (even the upper layer) to forcings. The ERF concept defined in AR5 allows rapid adjustments to perturbations, for all variables except for GMST or ocean temperature and sea ice cover. The ERF and RF values are significantly different for the anthropogenic aerosols, owing to their influence on clouds and on snow or ice cover. For other components that drive the Earth's energy balance, such as GHGs, ERF and RF are fairly similar, and RF may have comparable utility given that it requires fewer computational resources to calculate and is not affected by meteorological variability and hence can better isolate small forcings. In cases where RF and ERF differ substantially, ERF has been shown to be a better indicator of the GMST response and is therefore emphasized in AR5. {7.1, 8.1; Box 8.1}

AR4, N_2O has overtaken CFC-12 to become the third largest WMGHG contributor to RF. The RF from halocarbons is very similar to the value in AR4, with a reduced RF from CFCs but increases in many of their replacements. Four of the halocarbons (trichlorofluoromethane ($CFCl_3$, CFC-11), CFC-12, trichlorotrifluoroethane ($CF_2ClCFCl_2$, CFC-113) and chlorodifluoromethane (CHF_2Cl, HCFC-22) account for 85% of the total halocarbon RF. The former three compounds have declining RF over the last 5 years but are more than compensated for by the increased RF from HCFC-22. There is *high confidence* that the growth rate in RF from all WMGHG is weaker over the last decade than in the 1970s and 1980s owing to a slower increase in the non-CO_2 RF. {2.2.1, 8.3.2}

The short-lived GHGs ozone (O_3) and stratospheric water vapour also contribute to anthropogenic forcing. Observations indicate that O_3 *likely* increased at many undisturbed (background) locations through the 1990s. These increases have continued mainly over Asia (though

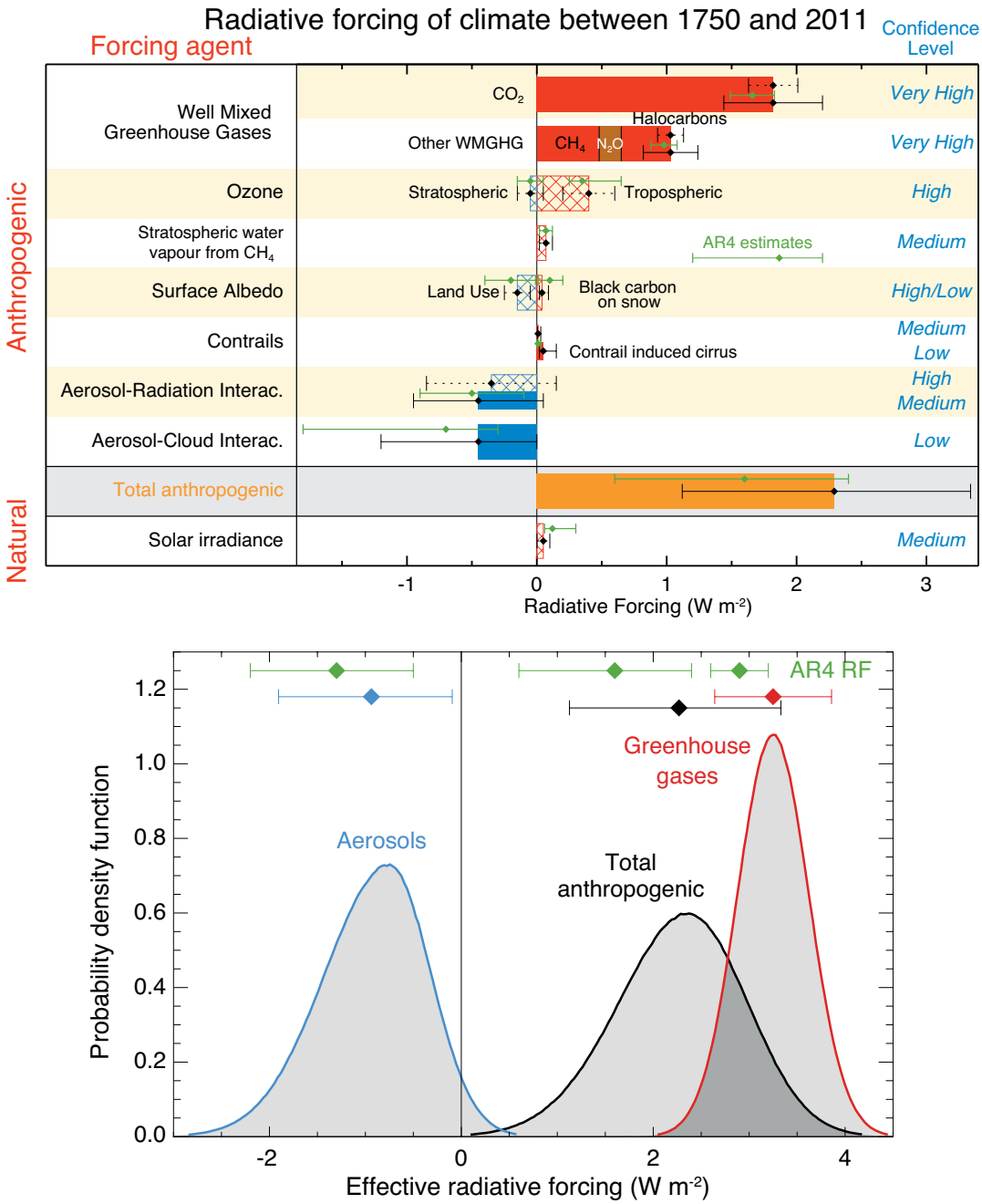

Figure TS.6 | Radiative forcing (RF) and Effective radiative forcing (ERF) of climate change during the Industrial Era. (Top) Forcing by concentration change between 1750 and 2011 with associated uncertainty range (solid bars are ERF, hatched bars are RF, green diamonds and associated uncertainties are for RF assessed in AR4). (Bottom) Probability density functions (PDFs) for the ERF, for the aerosol, greenhouse gas (GHG) and total. The green lines show the AR4 RF 90% confidence intervals and can be compared with the red, blue and black lines which show the AR5 ERF 90% confidence intervals (although RF and ERF differ, especially for aerosols). The ERF from surface albedo changes and combined contrails and contrail-induced cirrus is included in the total anthropogenic forcing, but not shown as a separate PDF. For some forcing mechanisms (ozone, land use, solar) the RF is assumed to be representative of the ERF but an additional uncertainty of 17% is added in quadrature to the RF uncertainty. {Figures 8.15, 8.16}

observations cover a limited area) and flattened over Europe during the last decade. The total RF due to changes in O_3 is 0.35 [0.15 to 0.55] W m^{-2} (*high confidence*), with RF due to tropospheric O_3 of 0.40 [0.20 to 0.60] W m^{-2} (*high confidence*) and due to stratospheric O_3 of –0.05 [–0.15 to +0.05] W m^{-2} (*high confidence*). O_3 is not emitted directly into the atmosphere; instead it is formed by photochemical reactions. In the troposphere these reactions involve precursor compounds that are emitted into the atmosphere from a variety of natural and anthropogenic sources. Tropospheric O_3 RF is largely attributed to increases in emissions of CH_4, carbon monoxide, volatile organics and nitrogen oxides, while stratospheric RF results primarily from O_3 depletion by anthropogenic halocarbons. However, there is now strong evidence for substantial links between the changes in tropospheric and stratospheric O_3 and a total O_3 RF of 0.50 [0.30 to 0.70] W m^{-2} is attributed to tropospheric O_3 precursor emissions and –0.15 [–0.30 to 0.00] W m^{-2} to O_3 depletion by halocarbons. There is strong evidence that tropospheric O_3 also has a detrimental impact on vegetation physiology, and therefore on its CO_2 uptake. This reduced uptake leads to an indirect increase in the atmospheric CO_2 concentration. Thus a fraction of the CO_2 RF should be attributed to ozone or its precursors rather than direct emission of CO_2, but there is a *low confidence* on the quantitative estimates. RF for stratospheric water vapour produced from CH_4 oxidation is 0.07 [0.02 to 0.12] W m^{-2}. Other changes in stratospheric water vapour, and all changes in water vapour in the troposphere, are regarded as a feedback rather than a forcing. {2.2.2, 8.1–8.3; FAQ 8.1}

TS.3.3 Radiative Forcing from Anthropogenic Aerosols

Anthropogenic aerosols are responsible for an RF of climate through multiple processes which can be grouped into two types: aerosol–radiation interactions (ari) and aerosol–cloud interactions (aci). There has been progress since AR4 on observing and modelling climate-relevant aerosol properties (including their size distribution, hygroscopicity, chemical composition, mixing state, optical and cloud nucleation properties) and their atmospheric distribution. Nevertheless, substantial uncertainties remain in assessments of long-term trends of global aerosol optical depth and other global properties of aerosols due to difficulties in measurement and lack of observations of some relevant parameters, high spatial and temporal variability and the relatively short observational records that exist. The anthropogenic RFari is given a best estimate of –0.35 [–0.85 to +0.15] W m^{-2} (*high confidence*) using evidence from aerosol models and some constraints from observations. The RFari is caused by multiple aerosol types (see Section TS3.6). The rapid adjustment to RFari leads to further negative forcing, in particular through cloud adjustments, and is attributable primarily to black carbon. As a consequence, the ERFari is more negative than the RFari (*low confidence*) and given a best estimate of –0.45 [–0.95 to +0.05] W m^{-2}. The assessment for RFari is less negative than reported in AR4 because of a re-evaluation of aerosol absorption. The uncertainty estimate is wider but more robust. {2.2.3, 7.3, 7.5.2}

Improved understanding of aerosol–cloud interactions has led to a reduction in the magnitude of many global aerosol–cloud forcings estimates. The total ERF due to aerosols (ERFari+aci, excluding the effect of absorbing aerosol on snow and ice) is assessed to be –0.9 [–1.9 to –0.1] W m^{-2} (*medium confidence*). This estimate encompasses all rapid adjustments, including changes to the cloud lifetime and aerosol microphysical effects on mixed-phase, ice and convective clouds. This range was obtained by giving equal weight to satellite-based studies and estimates from climate models. It is consistent with multiple lines of evidence suggesting less negative estimates for aerosol–cloud interactions than those discussed in AR4. {7.4, 7.5, 8.5}

The RF from black carbon (BC) on snow and ice is assessed to be 0.04 [0.02 to 0.09] W m^{-2} (*low confidence*). Unlike in the previous IPCC assessment, this estimate includes the effects on sea ice, accounts for more physical processes and incorporates evidence from both models and observations. This RF causes a two to four times larger GMST change per unit forcing than CO_2 primarily because all of the forcing energy is deposited directly into the cryosphere, whose evolution drives a positive albedo feedback on climate. This effect thus can represent a significant forcing mechanism in the Arctic and other snow- or ice-covered regions. {7.3, 7.5.2, 8.3.4, 8.5}

Despite the large uncertainty ranges on aerosol forcing, there is a *high confidence* that aerosols have offset a substantial portion of GHG forcing. Aerosol–cloud interactions can influence the character of individual storms, but evidence for a systematic aerosol effect on storm or precipitation intensity is more limited and ambiguous. {7.4, 7.6, 8.5}

TS.3.4 Radiative Forcing from Land Surface Changes and Contrails

There is robust evidence that anthropogenic land use changes such as deforestation have increased the land surface albedo, which leads to an RF of –0.15 [–0.25 to –0.05] W m^{-2}. There is still a large spread of quantitative estimates owing to different assumptions for the albedo of natural and managed surfaces (e.g., croplands, pastures). In addition, the time evolution of the land use change, and in particular how much was already completed in the reference year 1750, are still debated. Furthermore, land use change causes other modifications that are not radiative but impact the surface temperature, including modifications in the surface roughness, latent heat flux, river runoff and irrigation. These are more uncertain and they are difficult to quantify, but they tend to offset the impact of albedo changes at the global scale. As a consequence, there is low agreement on the sign of the net change in global mean temperature as a result of land use change. Land use change, and in particular deforestation, also has significant impacts on WMGHG concentrations. It contributes to the corresponding RF associated with CO_2 emissions or concentration changes. {8.3.5}

Persistent contrails from aviation contribute a positive RF of 0.01 [0.005 to 0.03] W m^{-2} (*medium confidence*) for year 2011, and the combined contrail and contrail-cirrus ERF from aviation is assessed to be 0.05 [0.02 to 0.15] W m^{-2} (*low confidence*). This forcing can be much larger regionally but there is now *medium confidence* that it does not produce observable regional effects on either the mean or diurnal range of surface temperature. {7.2.7}

TS.3.5 Radiative Forcing from Natural Drivers of Climate Change

Solar and volcanic forcings are the two dominant natural contributors to global climate change during the Industrial Era. Satellite observations

of total solar irradiance (TSI) changes since 1978 show quasi-periodic cyclical variation with a period of roughly 11 years. Longer term forcing is typically estimated by comparison of solar minima (during which variability is least). This gives an RF change of –0.04 [–0.08 to 0.00] W m^{-2} between the most recent (2008) minimum and the 1986 minimum. There is some diversity in the estimated trends of the composites of various satellite data, however. Secular trends of TSI before the start of satellite observations rely on a number of indirect proxies. The best estimate of RF from TSI changes over the industrial era is 0.05 [0.00 to 0.10] W m^{-2} (*medium confidence*), which includes greater RF up to around 1980 and then a small downward trend. This RF estimate is substantially smaller than the AR4 estimate due to the addition of the latest solar cycle and inconsistencies in how solar RF was estimated in earlier IPCC assessments. The recent solar minimum appears to have been unusually low and long-lasting and several projections indicate lower TSI for the forthcoming decades. However, current abilities to project solar irradiance are extremely limited so that there is *very low confidence* concerning future solar forcing. Nonetheless, there is a *high confidence* that 21st century solar forcing will be much smaller than the projected increased forcing due to WMGHGs. {5.2.1, 8.4.1; FAQ 5.1}

Changes in solar activity affect the cosmic ray flux impinging upon the Earth's atmosphere, which has been hypothesized to affect climate through changes in cloudiness. Cosmic rays enhance aerosol nucleation and thus may affect cloud condensation nuclei production in the free troposphere, but the effect is too weak to have any climatic influence during a solar cycle or over the last century (medium evidence, high agreement). No robust association between changes in cosmic rays and cloudiness has been identified. In the event that such an association existed, a mechanism other than cosmic ray–induced nucleation of new aerosol particles would be needed to explain it. {7.3, 7.4.6}

The RF of stratospheric volcanic aerosols is now well understood and there is a large RF for a few years after major volcanic eruptions (Box TS.5, Figure 1). Although volcanic eruptions inject both mineral particles and sulphate aerosol precursors into the atmosphere, it is the latter, because of their small size and long lifetimes, that are responsible for RF important for climate. The emissions of CO_2 from volcanic eruptions are at least 100 times smaller than anthropogenic emissions, and inconsequential for climate on century time scales. Large tropical volcanic eruptions have played an important role in driving annual to decadal scale climate change during the Industrial Era owing to their sometimes very large negative RF. There has not been any major volcanic eruption since Mt Pinatubo in 1991, which caused a 1-year RF of about –3.0 W m^{-2}, but several smaller eruptions have caused an RF averaged over the years 2008–2011 of –0.11 [–0.15 to –0.08] W m^{-2} (*high confidence*), twice as strong in magnitude compared to the 1999–2002 average. The smaller eruptions have led to better understanding of the dependence of RF on the amount of material from high-latitude injections as well as the time of the year when they take place. {5.2.1, 5.3.5, 8.4.2; Annex II}

TS.3.6 Synthesis of Forcings; Spatial and Temporal Evolution

A synthesis of the Industrial Era forcing finds that among the forcing agents, there is a *very high confidence* only for the WMGHG RF. Relative to AR4, the confidence level has been elevated for seven forcing agents owing to improved evidence and understanding. {8.5; Figure 8.14}

The time evolution of the total anthropogenic RF shows a nearly continuous increase from 1750, primarily since about 1860. The total anthropogenic RF increase rate since 1960 has been much greater than during earlier Industrial Era periods, driven primarily by the continuous increase in most WMGHG concentrations. There is still low agreement on the time evolution of the total aerosol ERF, which is the primary factor for the uncertainty in the total anthropogenic forcing. The fractional uncertainty in the total anthropogenic forcing decreases gradually after 1950 owing to the smaller offset of positive WMGHG forcing by negative aerosol forcing. There is robust evidence and high agreement that natural forcing is a small fraction of the WMGHG forcing. Natural forcing changes over the last 15 years have *likely* offset a substantial fraction (at least 30%) of the anthropogenic forcing increase during this period (Box TS.3). Forcing by CO_2 is the largest single contributor to the total forcing during the Industrial Era and from 1980–2011. Compared to the entire Industrial Era, the dominance of CO_2 forcing is larger for the 1980–2011 change with respect to other WMGHGs, and there is *high confidence* that the offset from aerosol forcing to WMGHG forcing during this period was much smaller than over the 1950–1980 period. {8.5.2}

Forcing can also be attributed to emissions rather than to the resulting concentration changes (Figure TS.7). Carbon dioxide is the largest single contributor to historical RF from either the perspective of changes in the atmospheric concentration of CO_2 or the impact of changes in net emissions of CO_2. The relative importance of other forcing agents can vary markedly with the perspective chosen, however. In particular, CH_4 emissions have a much larger forcing (about 1.0 W m^{-2} over the Industrial Era) than CH_4 concentration increases (about 0.5 W m^{-2}) due to several indirect effects through atmospheric chemistry. In addition, carbon monoxide emissions are *virtually certain* to cause a positive forcing, while emissions of reactive nitrogen oxides *likely* cause a net negative forcing but uncertainties are large. Emissions of ozone-depleting halocarbons *very likely* cause a net positive forcing as their direct radiative effect is larger than the impact of the stratospheric ozone depletion that they induce. Emissions of SO_2, organic carbon and ammonia cause a negative forcing, while emissions of black carbon lead to positive forcing via aerosol–radiation interactions. Note that mineral dust forcing may include a natural component or a climate feedback effect. {7.3, 7.5.2, 8.5.1}

Although the WMGHGs show a spatially fairly homogeneous forcing, other agents such as aerosols, ozone and land use changes are highly heterogeneous spatially. RFari showed maximum negative values over eastern North America and Europe during the early 20th century, with large negative values extending to East and Southeast Asia, South America and central Africa by 1980. Since then, however, the magnitude has decreased over eastern North America and Europe due to pollution control, and the peak negative forcing has shifted to South and East Asia primarily as a result of economic growth and the resulting increase in emissions in those areas. Total aerosol ERF shows similar behaviour for locations with maximum negative forcing, but also shows substantial positive forcing over some deserts and the Arctic. In contrast, the global mean whole atmosphere ozone forcing increased throughout

the 20th century, and has peak positive amplitudes around 15°N to 30°N but negative values over Antarctica. Negative land use forcing by albedo changes has been strongest in industrialized and biomass burning regions. The inhomogeneous nature of these forcings can cause them to have a substantially larger influence on the hydrologic cycle than an equivalent global mean homogeneous forcing. {8.3.5, 8.6}

Over the 21st century, anthropogenic RF is projected to increase under the Representative Concentration Pathways (RCPs; see Box TS.6). Simple model estimates of the RF resulting from the RCPs, which include WMGHG emissions spanning a broad range of possible futures, show anthropogenic RF relative to 1750 increasing to 3.0 to 4.8 W m^{-2} in 2050, and 2.7 to 8.4 W m^{-2} at 2100. In the near term, the RCPs are quite similar to one another (and emissions of near-term climate forcers do not span the literature range of possible futures), with RF at 2030 ranging only from 2.9 to 3.3 W m^{-2} (additional 2010 to 2030 RF of 0.7 to 1.1 W m^{-2}), but they show highly diverging values for the second half of the 21st century driven largely by CO_2. Results based on the RCP scenarios suggest only small changes in aerosol ERF between 2000 and 2030, followed by a strong reduction in the aerosols and a substantial weakening of the negative total aerosol ERF. Nitrate aerosols are an exception to this reduction, with a substantially increased negative forcing which is a robust feature among the few available models. The divergence across the RCPs indicates that, although a certain amount of future climate change is already 'in the system' due to the current radiative imbalance caused by historical emissions and the long lifetime of some atmospheric forcing agents, societal choices can still have a very large effect on future RF, and hence on climate change. {8.2, 8.5.3, 12.3; Figures 8.22, 12.4}

TS.3.7 Climate Feedbacks

Feedbacks will also play an important role in determining future climate change. Indeed, climate change may induce modification in the water, carbon and other biogeochemical cycles which may reinforce (positive feedback) or dampen (negative feedback) the expected

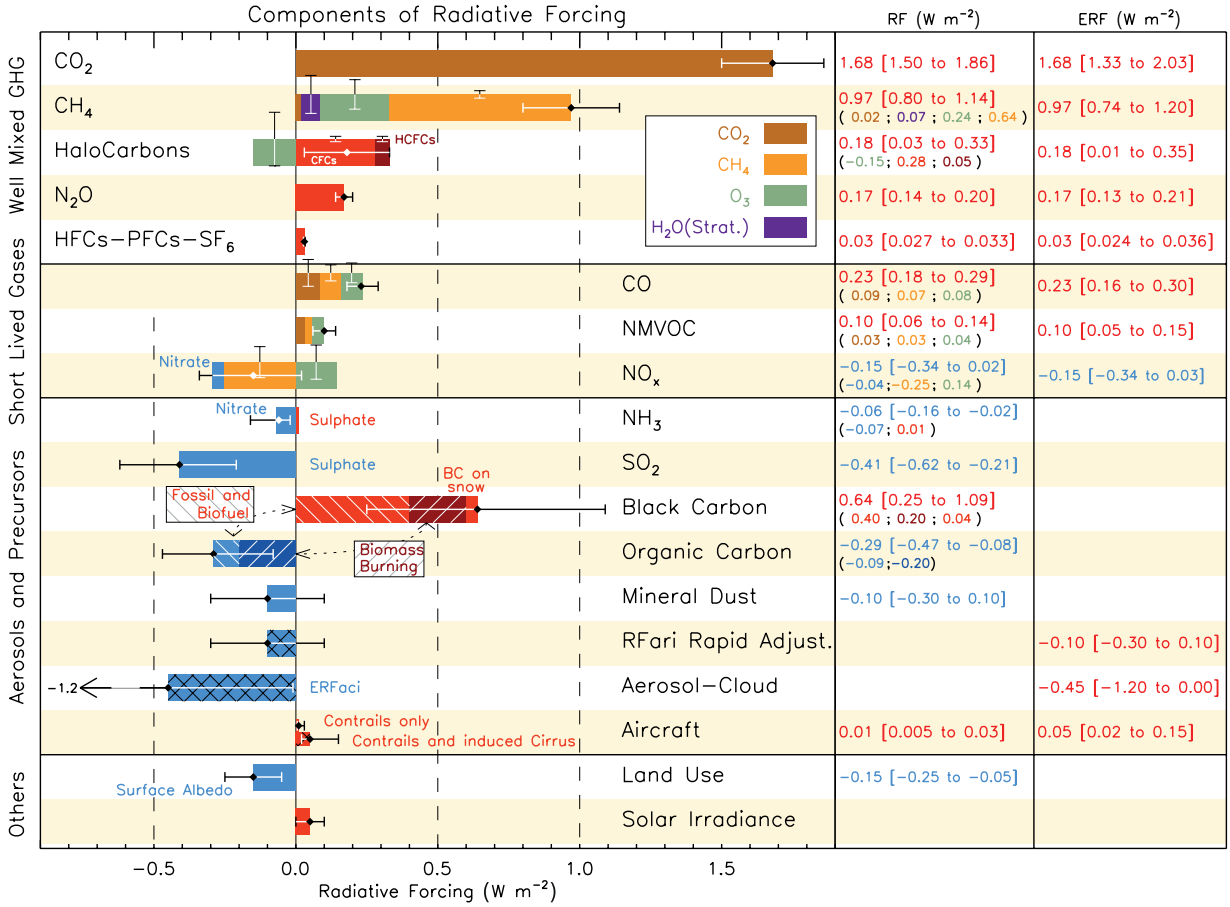

Figure TS.7 | Radiative forcing (RF) of climate change during the Industrial Era shown by emitted components from 1750 to 2011. The horizontal bars indicate the overall uncertainty, while the vertical bars are for the individual components (vertical bar lengths proportional to the relative uncertainty, with a total length equal to the bar width for a ±50% uncertainty). Best estimates for the totals and individual components (from left to right) of the response are given in the right column. Values are RF except for the effective radiative forcing (ERF) due to aerosol–cloud interactions (ERFaci) and rapid adjustment associated with the RF due to aerosol-radiation interaction (RFari Rapid Adjust.). Note that the total RF due to aerosol-radiation interaction (–0.35 Wm^{-2}) is slightly different from the sum of the RF of the individual components (–0.33 Wm^{-2}). The total RF due to aerosol-radiation interaction is the basis for Figure SPM.5. Secondary organic aerosol has not been included since the formation depends on a variety of factors not currently sufficiently quantified. The ERF of contrails includes contrail induced cirrus. Combining ERFaci –0.45 [–1.2 to 0.0] Wm^{-2} and rapid adjustment of ari –0.1 [–0.3 to +0.1] Wm^{-2} results in an integrated component of adjustment due to aerosols of –0.55 [–1.33 to –0.06] Wm^{-2}. CFCs = chlorofluorocarbons, HCFCs = hydrochlorofluorocarbons, HFCs = hydrofluorocarbons, PFCs = perfluorocarbons, NMVOC = Non-Methane Volatile Organic Compounds, BC = black carbon. Further detail regarding the related Figure SPM.5 is given in the TS Supplementary Material. {Figure 8.17}

temperature increase. Snow and ice albedo feedbacks are known to be positive. The combined water vapour and lapse rate feedback is *extremely likely* to be positive and now fairly well quantified, while cloud feedbacks continue to have larger uncertainties (see TFE.6). In addition, the new Coupled Model Intercomparison Project Phase 5 (CMIP5) models consistently estimate a positive carbon-cycle feedback, that is, reduced natural CO_2 sinks in response to future climate change. In particular, carbon-cycle feedbacks in the oceans are positive in the models. Carbon sinks in tropical land ecosystems are less consistent, and may be susceptible to climate change via processes such as drought and fire that are sometimes not yet fully represented. A key update since AR4 is the introduction of nutrient dynamics in some of the CMIP5 land carbon models, in particular the limitations on plant growth imposed by nitrogen availability. The net effect of accounting for the nitrogen cycle is a smaller projected land sink for a given trajectory of anthropogenic CO_2 emissions (see TFE.7). {6.4, Box 6.1, 7.2}

Models and ecosystem warming experiments show high agreement that wetland CH_4 emissions will increase per unit area in a warmer climate, but wetland areal extent may increase or decrease depending on regional changes in temperature and precipitation affecting wetland hydrology, so that there is *low confidence* in quantitative projections of wetland CH_4 emissions. Reservoirs of carbon in hydrates and permafrost are very large, and thus could potentially act as very powerful feedbacks. Although poorly constrained, the 21st century global release of CH_4 from hydrates to the atmosphere is *likely* to be low due to the under-saturated state of the ocean, long ventilation time of the ocean and slow propagation of warming through the seafloor. There is *high confidence* that release of carbon from thawing permafrost provides a positive feedback, but there is *low confidence* in quantitative projections of its strength. {6.4.7}

Aerosol-climate feedbacks occur mainly through changes in the source strength of natural aerosols or changes in the sink efficiency of natural and anthropogenic aerosols; a limited number of modelling studies have assessed the magnitude of this feedback to be small with a *low confidence*. There is *medium confidence* for a weak feedback (of uncertain sign) involving dimethylsulphide, cloud condensation nuclei and cloud albedo due to a weak sensitivity of cloud condensation nuclei population to changes in dimethylsulphide emissions. {7.3.5}

TS.3.8 Emission Metrics

Different metrics can be used to quantify and communicate the relative and absolute contributions to climate change of emissions of different substances, and of emissions from regions/countries or sources/sectors. Up to AR4, the most common metric has been the Global Warming Potential (GWP) that integrates RF out to a particular time horizon. This metric thus accounts for the radiative efficiencies of the various substances, and their lifetimes in the atmosphere, and gives values relative to those for the reference gas CO_2. There is now increasing focus on the Global Temperature change Potential (GTP), which is based on the change in GMST at a chosen point in time, again relative to that caused by the reference gas CO_2, and thus accounts for climate response along with radiative efficiencies and atmospheric lifetimes. Both the GWP and the GTP use a time horizon (Figure TS.8 top), the choice of which is subjective and context dependent. In general, GWPs for near-term climate forcers are higher than GTPs due to the equal time weighting in the integrated forcing used in the GWP. Hence the choice of metric can greatly affect the relative importance of near-term climate forcers and WMGHGs, as can the choice of time horizon. Analysis of the impact of current emissions (1-year pulse of emissions) shows that near-term climate forcers, such as black carbon, sulphur dioxide or CH_4, can have contributions comparable to that of CO_2 for short time horizons (of either the same or opposite sign), but their impacts become progressively less for longer time horizons over which emissions of CO_2 dominate (Figure TS.8 top). {8.7}

A large number of other metrics may be defined down the driver–response–impact chain. No single metric can accurately compare all consequences (i.e., responses in climate parameters over time) of different emissions, and a metric that establishes equivalence with regard to one effect will not give equivalence with regard to other effects. The choice of metric therefore depends strongly on the particular consequence one wants to evaluate. It is important to note that the metrics do not define policies or goals, but facilitate analysis and implementation of multi-component policies to meet particular goals. All choices of metric contain implicit value-related judgements such as type of effect considered and weighting of effects over time. Whereas GWP integrates the effects up to a chosen time horizon (i.e., giving equal weight to all times up to the horizon and zero weight thereafter), the GTP gives the temperature just for one chosen year with no weight on years before or after. {8.7}

The GWP and GTP have limitations and suffer from inconsistencies related to the treatment of indirect effects and feedbacks, for instance, if climate–carbon feedbacks are included for the reference gas CO_2 but not for the non-CO_2 gases. The uncertainty in the GWP increases with time horizon, and for the 100-year GWP of WMGHGs the uncertainty can be as large as ±40%. Several studies also point out that this metric is not well suited for policies with a maximum temperature target. Uncertainties in GTP also increase with time as they arise from the same factors contributing to GWP uncertainties along with additional contributions from it being further down the driver–response–impact chain and including climate response. The GTP metric is better suited to target-based policies, but is again not appropriate for every goal. Updated metric values accounting for changes in knowledge of lifetimes and radiative efficiencies and for climate–carbon feedbacks are now available. {8.7, Table 8.7, Table 8.A.1, Chapter 8 Supplementary Material Table 8.SM.16}

With these emission metrics, the climate impact of past or current emissions attributable to various activities can be assessed. Such activity-based accounting can provide additional policy-relevant information, as these activities are more directly affected by particular societal choices than overall emissions. A single year's worth of emissions (a pulse) is often used to quantify the impact on future climate. From this perspective and with the absolute GTP metric used to illustrate the results, energy and industry have the largest contributions to warming over the next 50 to 100 years (Figure TS.8, bottom). Household fossil and biofuel, biomass burning and on-road transportation are also relatively large contributors to warming over these time scales, while current emissions from sectors that emit large amounts of CH_4 (animal husbandry, waste/landfills and agriculture) are also important over