

## 8.1 Radiative Forcing

There are a variety of ways to examine how various drivers contribute to climate change. In principle, observations of the climate response to a single factor could directly show the impact of that factor, or climate models could be used to study the impact of any single factor. In practice, however, it is usually difficult to find measurements that are influenced by only a single cause, and it is computationally prohibitive to simulate the response to every individual factor of interest. Hence various metrics intermediate between cause and effect are used to provide estimates of the climate impact of individual factors, with applications both in science and policy. Radiative forcing (RF) is one of the most widely used metrics, with most other metrics based on RF. In this chapter, we discuss RF from natural and anthropogenic components during the industrial period, presenting values for 2011 relative to 1750 unless otherwise stated, and projected values through 2100 (see also Annex II). In this section, we present the various definitions of RF used in this chapter, and discuss the utility and limitations of RF. These definitions are used in the subsequent sections quantifying the RF due to specific anthropogenic (Section 8.3) and natural (Section 8.4) causes and integrating RF due to all causes (Sections 8.5 and 8.6). Atmospheric chemistry relevant for RF is discussed in Section 8.2 and used throughout the chapter. Emission metrics using RF that are designed to facilitate rapid evaluation and comparison of the climate effects of emissions are discussed in Section 8.7.

### 8.1.1 The Radiative Forcing Concept

RF is the net change in the energy balance of the Earth system due to some imposed perturbation. It is usually expressed in watts per square meter averaged over a particular period of time and quantifies the energy imbalance that occurs when the imposed change takes place. Though usually difficult to observe, calculated RF provides a simple quantitative basis for comparing some aspects of the potential climate response to different imposed agents, especially global mean temperature, and hence is widely used in the scientific community. Forcing is often presented as the value due to changes between two particular times, such as pre-industrial to present-day, while its time evolution provides a more complete picture.

#### 8.1.1.1 Defining Radiative Forcing

Alternative definitions of RF have been developed, each with its own advantages and limitations. The instantaneous RF refers to an instantaneous change in net (down minus up) radiative flux (shortwave plus longwave; in  $\text{W m}^{-2}$ ) due to an imposed change. This forcing is usually defined in terms of flux changes at the top of the atmosphere (TOA) or at the climatological tropopause, with the latter being a better indicator of the global mean surface temperature response in cases when they differ.

Climate change takes place when the system responds in order to counteract the flux changes, and all such responses are explicitly

excluded from this definition of forcing. The assumed relation between a sustained RF and the equilibrium global mean surface temperature response ( $\Delta T$ ) is  $\Delta T = \lambda \text{RF}$  where  $\lambda$  is the climate sensitivity parameter. The relationship between RF and  $\Delta T$  is an expression of the energy balance of the climate system and a simple reminder that the steady-state global mean climate response to a given forcing is determined both by the forcing and the responses inherent in  $\lambda$ .

Implicit in the concept of RF is the proposition that the change in net irradiance in response to the imposed forcing alone can be separated from all subsequent responses to the forcing. These are not in fact always clearly separable and thus some ambiguity exists in what may be considered a forcing versus what is part of the climate response.

In both the Third Assessment Report (TAR) and AR4, the term radiative forcing (RF, also called stratospherically adjusted RF, as distinct from instantaneous RF) was defined as the change in net irradiance at the tropopause after allowing for stratospheric temperatures to readjust to radiative equilibrium, while holding surface and tropospheric temperatures and state variables such as water vapour and cloud cover fixed at the unperturbed values<sup>8</sup>. RF is generally more indicative of the surface and tropospheric temperature responses than instantaneous RF, especially for agents such as carbon dioxide ( $\text{CO}_2$ ) or ozone ( $\text{O}_3$ ) change that substantially alter stratospheric temperatures. To be consistent with TAR and AR4, RF is hereafter taken to mean the stratospherically adjusted RF.

#### 8.1.1.2 Defining Effective Radiative Forcing

For many forcing agents the RF gives a very useful and appropriate way to compare the relative importance of their potential climate effect. Instantaneous RF or RF is not an accurate indicator of the temperature response for all forcing agents, however. Rapid adjustments in the troposphere can either enhance or reduce the flux perturbations, leading to substantial differences in the forcing driving long-term climate change. In much the same way that allowing for the relatively rapid adjustment of stratospheric temperatures provides a more useful characterization of the forcing due to stratospheric constituent changes, inclusion of rapid tropospheric adjustments has the potential to provide more useful characterization for drivers in the troposphere (see also Section 7.1.3).

Many of the rapid adjustments affect clouds and are not readily included into the RF concept. For example, for aerosols, especially absorbing ones, changes in the temperature distribution above the surface occur due to a variety of effects, including cloud response to changing atmospheric stability (Hansen et al., 2005; see Section 7.3.4.2) and cloud absorption effects (Jacobson, 2012), which affect fluxes but are not strictly part of RF. Similar adjustments take place for many forcings, including  $\text{CO}_2$  (see Section 7.2.5.6).

Aerosols also alter cloud properties via microphysical interactions leading to indirect forcings (referred to as aerosol–cloud interactions;

<sup>8</sup> Tropospheric variables were fixed except for the impact of aerosols on cloud albedo due to changes in droplet size with constant cloud liquid water which was considered an RF in AR4 but is part of ERF in AR5.

see Section 7.4). Although these adjustments are complex and not fully quantified, they occur both on the microphysical scale of the cloud particles as well as on a more macroscopic scale involving whole cloud systems (e.g., Shine et al., 2003; Penner et al., 2006; Quaas et al., 2009). A portion of these adjustments occurs over a short period, on cloud life cycle time scales, and is not part of a feedback arising from the surface temperature changes. Previously these type of adjustments were sometimes termed ‘fast feedbacks’ (e.g., Gregory et al., 2004; Hansen et al., 2005), whereas in AR5 they are denoted ‘rapid adjustments’ to emphasize their distinction from feedbacks involving surface temperature changes. Atmospheric chemistry responses have typically been included under the RF framework, and hence could also be included in a forcing encompassing rapid adjustments, which is important when evaluating forcing attributable to emissions changes (Section 8.1.2) and in the calculation of emission metrics (Section 8.7).

Studies have demonstrated the utility of including rapid adjustment in comparison of forcing agents, especially in allowing quantification of forcing due to aerosol-induced changes in clouds (e.g., effects previously denoted as cloud lifetime or semi-direct effects; see Figure 7.3) that are not amenable to characterization by RF (e.g., Rotstayn and Penner, 2001; Shine et al., 2003; Hansen et al., 2005; Lohmann et al., 2010; Ban-Weiss et al., 2012). Several measures of forcing have been introduced that include rapid adjustments. We term a forcing that accounts for rapid adjustments the effective radiative forcing (ERF). Conceptually, ERF represents the change in net TOA downward radiative flux after allowing for atmospheric temperatures, water vapour and clouds to adjust, but with global mean surface temperature or a portion of surface conditions unchanged. The primary methods in use for such calculations are (1) fixing sea surface temperatures (SSTs) and sea ice cover at climatological values while allowing all other parts of the system to respond until reaching steady state (e.g., Hansen et al., 2005) or (2) analyzing the transient global mean surface temperature response to an instantaneous perturbation and using the regression of the response extrapolated back to the start of the simulation to derive

the initial ERF (Gregory et al., 2004; Gregory and Webb, 2008). The ERF calculated using the regression technique has an uncertainty of about 10% (for the 5 to 95% confidence interval) for a single  $4 \times \text{CO}_2$  simulation (ERF  $\sim 7 \text{ W m}^{-2}$ ) due to internal variability in the transient climate (Andrews et al., 2012a), while given a similar length simulation the uncertainty due to internal variability in ERF calculated using the fixed-SST technique is much smaller and hence the latter may be more suitable for very small forcings. Analysis of both techniques shows that the fixed-SST method yields a smaller spread across models, even in calculations neglecting the uncertainty in the regression fitting procedure (Andrews et al., 2012a). As a portion of land area responses are included in the fixed-SST technique, however, that ERF is slightly less than it would be with surface temperature held fixed everywhere. It is possible to adjust for this in the global mean forcing, though we do not include such a correction here as we examine regional as well as global ERF, but the land response will also introduce artificial gradients in land–sea temperatures that could cause small local climate responses. In contrast, there is no global mean temperature response included in the regression method. Despite the low bias in fixed-SST ERF due to land responses, results from a multi-model analysis of the forcing due to  $\text{CO}_2$  are 7% greater using this method than using the regression technique (Andrews et al., 2012a) though this is within the uncertainty range of the calculations. Although each technique has advantages, forcing diagnosed using the fixed-SST method is available for many more forcing agents in the current generation of climate models than forcing diagnosed using the regression method. Hence for practical purposes, ERF is hereafter used for results from the fixed-SST technique unless otherwise stated (see also Box 8.1).

The conceptual relation between instantaneous RF, RF and ERF is illustrated in Figure 8.1. It implies the adjustments to the instantaneous RF involve effects of processes that occur more rapidly than the time scale of the response of the global mean surface temperature to the forcing. However, there is no *a priori* time scale defined for adjustments to be rapid with the fixed-SST method. The majority take place on time scales

### Box 8.1 | Definition of Radiative Forcing and Effective Radiative Forcing

The two most commonly used measures of radiative forcing in this chapter are the radiative forcing (RF) and the effective radiative forcing (ERF). RF is defined, as it was in AR4, as the change in net downward radiative flux at the tropopause after allowing for stratospheric temperatures to readjust to radiative equilibrium, while holding surface and tropospheric temperatures and state variables such as water vapor and cloud cover fixed at the unperturbed values.

ERF is the change in net TOA downward radiative flux after allowing for atmospheric temperatures, water vapour and clouds to adjust, but with surface temperature or a portion of surface conditions unchanged. Although there are multiple methods to calculate ERF, we take ERF to mean the method in which sea surface temperatures and sea ice cover are fixed at climatological values unless otherwise specified. Land surface properties (temperature, snow and ice cover and vegetation) are allowed to adjust in this method. Hence ERF includes both the effects of the forcing agent itself and the rapid adjustments to that agent (as does RF, though stratospheric temperature is the only adjustment for the latter). In the case of aerosols, the rapid adjustments of clouds encompass effects that have been referred to as indirect or semi-direct forcings (see Figure 7.3 and Section 7.5), with some of these same cloud responses also taking place for other forcing agents (see Section 7.2). Calculation of ERF requires longer simulations with more complex models than calculation of RF, but the inclusion of the additional rapid adjustments makes ERF a better indicator of the eventual global mean temperature response, especially for aerosols. When forcing is attributed to emissions or used for calculation of emission metrics, additional responses including atmospheric chemistry and the carbon cycle are also included in both RF and ERF (see Section 8.1.2). The general term *forcing* is used to refer to both RF and ERF.

## Frequently Asked Questions

**FAQ 8.1 | How Important Is Water Vapour to Climate Change?**

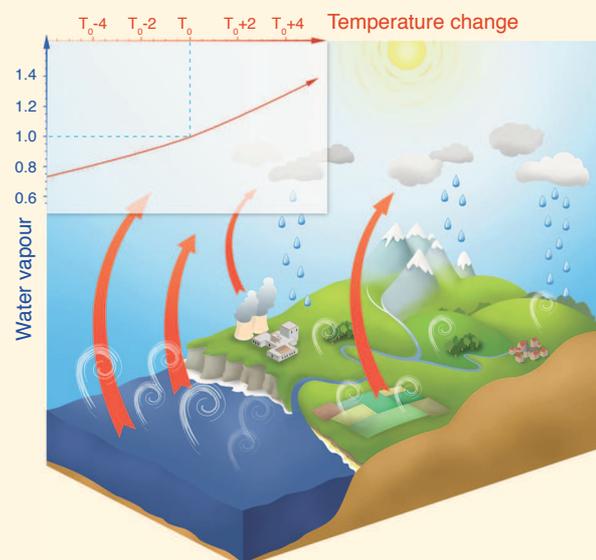
As the largest contributor to the natural greenhouse effect, water vapour plays an essential role in the Earth's climate. However, the amount of water vapour in the atmosphere is controlled mostly by air temperature, rather than by emissions. For that reason, scientists consider it a feedback agent, rather than a forcing to climate change. Anthropogenic emissions of water vapour through irrigation or power plant cooling have a negligible impact on the global climate.

Water vapour is the primary greenhouse gas in the Earth's atmosphere. The contribution of water vapour to the natural greenhouse effect relative to that of carbon dioxide (CO<sub>2</sub>) depends on the accounting method, but can be considered to be approximately two to three times greater. Additional water vapour is injected into the atmosphere from anthropogenic activities, mostly through increased evaporation from irrigated crops, but also through power plant cooling, and marginally through the combustion of fossil fuel. One may therefore question why there is so much focus on CO<sub>2</sub>, and not on water vapour, as a forcing to climate change.

Water vapour behaves differently from CO<sub>2</sub> in one fundamental way: it can condense and precipitate. When air with high humidity cools, some of the vapour condenses into water droplets or ice particles and precipitates. The typical residence time of water vapour in the atmosphere is ten days. The flux of water vapour into the atmosphere from anthropogenic sources is considerably less than from 'natural' evaporation. Therefore, it has a negligible impact on overall concentrations, and does not contribute significantly to the long-term greenhouse effect. This is the main reason why tropospheric water vapour (typically below 10 km altitude) is not considered to be an anthropogenic gas contributing to radiative forcing.

Anthropogenic emissions do have a significant impact on water vapour in the stratosphere, which is the part of the atmosphere above about 10 km. Increased concentrations of methane (CH<sub>4</sub>) due to human activities lead to an additional source of water, through oxidation, which partly explains the observed changes in that atmospheric layer. That stratospheric water change has a radiative impact, is considered a forcing, and can be evaluated. Stratospheric concentrations of water have varied significantly in past decades. The full extent of these variations is not well understood and is probably less a forcing than a feedback process added to natural variability. The contribution of stratospheric water vapour to warming, both forcing and feedback, is much smaller than from CH<sub>4</sub> or CO<sub>2</sub>.

The maximum amount of water vapour in the air is controlled by temperature. A typical column of air extending from the surface to the stratosphere in polar regions may contain only a few kilograms of water vapour per square metre, while a similar column of air in the tropics may contain up to 70 kg. With every extra degree of air temperature, the atmosphere can retain around 7% more water vapour (see upper-left insert in the FAQ 8.1, Figure 1). This increase in concentration amplifies the greenhouse effect, and therefore leads to more warming. This process, referred to as the water vapour feedback, is well understood and quantified. It occurs in all models used to estimate climate change, where its strength is consistent with observations. Although an increase in atmospheric water vapour has been observed, this change is recognized as a climate feedback (from increased atmospheric temperature) and should not be interpreted as a radiative forcing from anthropogenic emissions. *(continued on next page)*



**FAQ 8.1, Figure 1 |** Illustration of the water cycle and its interaction with the greenhouse effect. The upper-left insert indicates the relative increase of potential water vapour content in the air with an increase of temperature (roughly 7% per degree). The white curls illustrate evaporation, which is compensated by precipitation to close the water budget. The red arrows illustrate the outgoing infrared radiation that is partly absorbed by water vapour and other gases, a process that is one component of the greenhouse effect. The stratospheric processes are not included in this figure.

## FAQ 8.1 (continued)

Currently, water vapour has the largest greenhouse effect in the Earth's atmosphere. However, other greenhouse gases, primarily CO<sub>2</sub>, are necessary to sustain the presence of water vapour in the atmosphere. Indeed, if these other gases were removed from the atmosphere, its temperature would drop sufficiently to induce a decrease of water vapour, leading to a runaway drop of the greenhouse effect that would plunge the Earth into a frozen state. So greenhouse gases other than water vapour provide the temperature structure that sustains current levels of atmospheric water vapour. Therefore, although CO<sub>2</sub> is the main anthropogenic control knob on climate, water vapour is a strong and fast feedback that amplifies any initial forcing by a typical factor between two and three. Water vapour is not a significant initial forcing, but is nevertheless a fundamental agent of climate change.

of seasons or less, but there is a spectrum of adjustment times. Changes in land ice and snow cover, for instance, may take place over many years. The ERF thus represents that part of the instantaneous RF that is maintained over long time scales and more directly contributes to the steady-state climate response. The RF can be considered a more limited version of ERF. Because the atmospheric temperature has been allowed to adjust, ERF would be nearly identical if calculated at the tropopause instead of the TOA for tropospheric forcing agents, as would RF. Recent work has noted likely advantages of the ERF framework for understanding model responses to CO<sub>2</sub> as well as to more complex forcing agents (see Section 7.2.5.6).

The climate sensitivity parameter  $\lambda$ , derived with respect to RF can vary substantially across different forcing agents (Forster et al., 2007). The response to RF from a particular agent relative to the response to RF from CO<sub>2</sub> has been termed the *efficacy* (Hansen et al., 2005). By including many of the rapid adjustments that differ across forcing agents, the ERF concept includes much of their relative efficacy and therefore leads to more uniform climate sensitivity across agents. For example, the influence of clouds on the interaction of aerosols with sunlight and the effect of aerosol heating on cloud formation can lead to very large differences in the response per unit RF from black carbon (BC) located at different altitudes, but the response per unit ERF is nearly uniform with altitude (Hansen et al., 2005; Ming et al., 2010; Ban-Weiss et al., 2012). Hence as we use ERF in this chapter when it differs significantly from RF, efficacy is not used hereinafter. For inhomogeneous forcings, we note that the climate sensitivity parameter may also depend on the horizontal forcing distribution, especially with latitude (Shindell and Faluvegi, 2009; Section 8.6.2).

A combination of RF and ERF will be used in this chapter with RF provided to keep consistency with TAR and AR4, and ERF used to allow quantification of more complex forcing agents and, in some cases, provide a more useful metric than RF.

### 8.1.1.3 Limitations of Radiative Forcing

Both the RF and ERF concepts have strengths and weaknesses in addition to those discussed previously. Dedicated climate model simulations that are required to diagnose the ERF can be more computationally demanding than those for instantaneous RF or RF because many years are required to reduce the influence of climate variability. The presence of meteorological variability can also make it difficult to

isolate the ERF of small forcings that are easily isolated in the pair of radiative transfer calculations performed for RF (Figure 8.1). For RF, on the other hand, a definition of the tropopause is required, which can be ambiguous.

In many cases, however, ERF and RF are nearly equal. Analysis of 11 models from the current Coupled Model Intercomparison Project Phase 5 (CMIP5) generation finds that the rapid adjustments to CO<sub>2</sub> cause fixed-SST-based ERF to be 2% less than RF, with an intermodel standard deviation of 7% (Vial et al., 2013). This is consistent with an earlier study of six GCMs that found a substantial inter-model variation in the rapid tropospheric adjustment to CO<sub>2</sub> using regression analysis in slab ocean models, though the ensemble mean adjustment was less than 5% (Andrews and Forster, 2008). Part of the large uncertainty range arises from the greater noise inherent in regression analyses of single runs in comparison with fixed-SST experiments. Using fixed-SST simulations, Hansen et al. (2005) found that ERF is virtually identical to RF for increased CO<sub>2</sub>, tropospheric ozone and solar irradiance, and within 6% for methane (CH<sub>4</sub>), nitrous oxide (N<sub>2</sub>O), stratospheric aerosols and for the aerosol–radiation interaction of reflective aerosols. Shindell et al. (2013b) also found that RF and ERF are statistically equal for tropospheric ozone. Lohmann et al. (2010) report a small increase in the forcing from CO<sub>2</sub> using ERF instead of RF based on the fixed-SST technique, while finding no substantial difference for CH<sub>4</sub>, RF due to aerosol–radiation interactions or aerosol effects on cloud albedo. In the fixed-SST simulations of Hansen et al. (2005), ERF was about 20% less than RF for the atmospheric effects of BC aerosols (not including microphysical aerosol–cloud interactions), and nearly 300% greater for the forcing due to BC snow albedo forcing (Hansen et al., 2007). ERF was slightly greater than RF for stratospheric ozone in Hansen et al. (2005), but the opposite is true for more recent analyses (Shindell et al., 2013b), and hence it seems most appropriate at present to use RF for this small forcing. The various studies demonstrate that RF provides a good estimate of ERF in most cases, as the differences are very small, with the notable exceptions of BC-related forcings (Bond et al., 2013). ERF provides better characterization of those effects, as well as allowing quantification of a broader range of effects including all aerosol–cloud interactions. Hence while RF and ERF are generally quite similar for WMGHGs, ERF typically provides a more useful indication of climate response for near-term climate forcings (see Box 8.2). As the rapid adjustments included in ERF differ in strength across climate models, the uncertainty range for ERF estimates tends to be larger than the range for RF estimates.

### Box 8.2 | Grouping Forcing Compounds by Common Properties

As many compounds cause RF when their atmospheric concentration is changed, it can be useful to refer to groups of compounds with similar properties. Here we discuss two primary groupings: well-mixed greenhouse gases (WMGHGs) and near-term climate forcers (NTCFs).

We define as 'well-mixed' those greenhouse gases that are sufficiently mixed throughout the troposphere that concentration measurements from a few remote surface sites can characterize the climate-relevant atmospheric burden; although these gases may still have local variation near sources and sinks and even small hemispheric gradients. Global forcing per unit emission and emission metrics for these gases thus do not depend on the geographic location of the emission, and forcing calculations can assume even horizontal distributions. These gases, or a subset of them, have sometimes been referred to as 'long-lived greenhouse gases' as they are well mixed because their atmospheric lifetimes are much greater than the time scale of a few years for atmospheric mixing, but the physical property that causes the aforementioned common characteristics is more directly associated with their mixing within the atmosphere. WMGHGs include CO<sub>2</sub>, N<sub>2</sub>O, CH<sub>4</sub>, SF<sub>6</sub>, and many halogenated species. Conversely, ozone is not a WMGHG.

We define 'near-term climate forcers' (NTCFs) as those compounds whose impact on climate occurs primarily within the first decade after their emission. This set of compounds is composed primarily of those with short lifetimes in the atmosphere compared to WMGHGs, and has been sometimes referred to as short-lived climate forcers or short-lived climate pollutants. However, the common property that is of greatest interest to a climate assessment is the time scale over which their impact on climate is felt. This set of compounds includes methane, which is also a WMGHG, as well as ozone and aerosols, or their precursors, and some halogenated species that are not WMGHGs. These compounds do not accumulate in the atmosphere at decadal to centennial time scales, and so their effect on climate is predominantly in the near term following their emission.

Whereas the global mean ERF provides a useful indication of the eventual change in global mean surface temperature, it does not reflect regional climate changes. This is true for all forcing agents, but is especially the case for the inhomogeneously distributed forcings because they activate climate feedbacks based on their regional distribution. For example, forcings over Northern Hemisphere (NH) middle and high latitudes induce snow and ice albedo feedbacks more than forcings at lower latitudes or in the Southern Hemisphere (SH) (e.g., Shindell and Faluvegi, 2009).

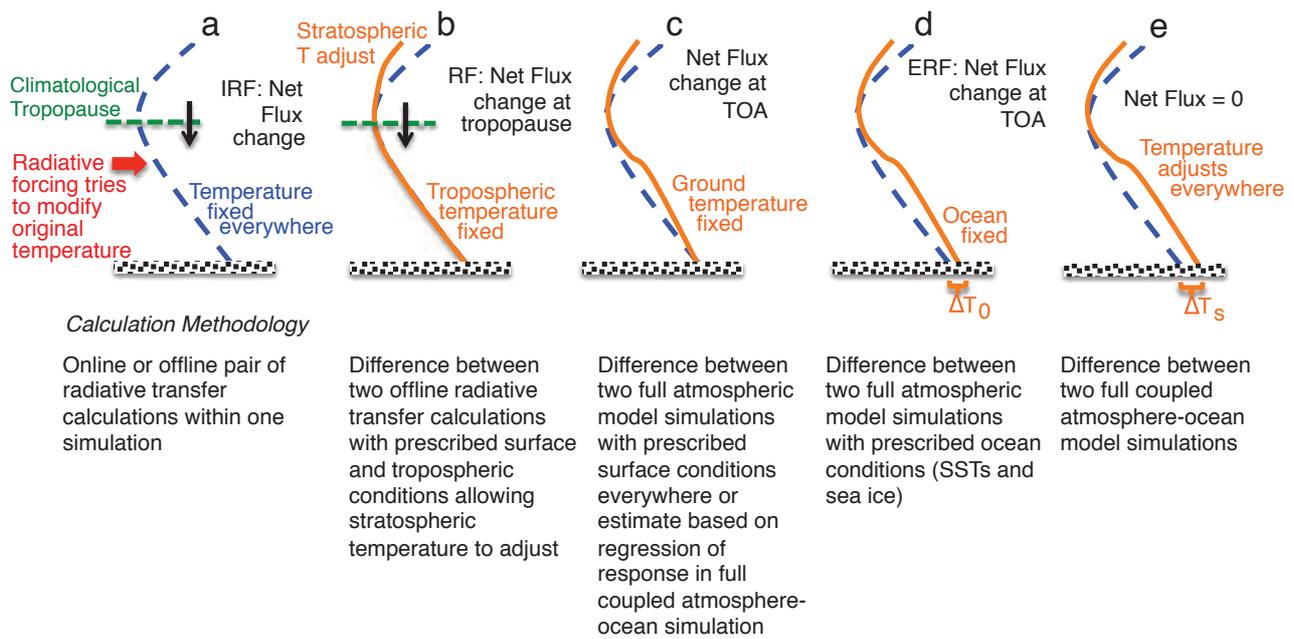
In the case of agents that strongly absorb incoming solar radiation (such as BC, and to a lesser extent organic carbon (OC) and ozone) the TOA forcing provides little indication of the change in solar radiation reaching the surface which can force local changes in evaporation and alter regional and general circulation patterns (e.g., Ramanathan and Carmichael, 2008; Wang et al., 2009). Hence the forcing at the surface, or the atmospheric heating, defined as the difference between surface and tropopause/TOA forcing, might also be useful metrics. Global mean precipitation changes can be related separately to ERF within the atmosphere and to a slower response to global mean temperature changes (Andrews et al., 2010; Ming et al., 2010; Ban-Weiss et al., 2012). Relationships between surface forcing and localized aspects of climate response have not yet been clearly quantified, however.

In general, most widely used definitions of forcing and most forcing-based metrics are intended to be proportional to the eventual temperature response, and most analyses to date have explored the global mean temperature response only. These metrics do not explicitly include impacts such as changes in precipitation, surface sunlight available for photosynthesis, extreme events, and so forth, or regional

temperatures, which can differ greatly from the global mean. Hence although they are quite useful for understanding the factors driving global mean temperature change, they provide only an imperfect and limited perspective on the factors driving broader climate change. In addition, a metric based solely on radiative perturbations does not allow comparison of non-RFs, such as effects of land cover change on evapotranspiration or physiological impacts of CO<sub>2</sub> and O<sub>3</sub> except where these cause further impacts on radiation such as through cloud cover changes (e.g., Andrews et al., 2012b).

#### 8.1.2 Calculation of Radiative Forcing due to Concentration or Emission Changes

Analysis of forcing due to observed or modelled concentration changes between pre-industrial, defined here as 1750, and a chosen later year provides an indication of the importance of different forcing agents to climate change during that period. Such analyses have been a mainstay of climate assessments. This perspective has the advantage that observational data are available to accurately quantify the concentration changes for several of the largest forcing components. Atmospheric concentration changes, however, are the net result of variations in emissions of multiple compounds and any climate changes that have influenced processes such as wet removal, atmospheric chemistry or the carbon cycle. Characterizing forcing according to *concentration* changes thus mixes multiple root causes along with climate feedbacks. Policy decisions are better informed by analysis of forcing attributable to *emissions*, which the IPCC first presented in AR4. These analyses can be applied to historical emissions changes in a 'backward-looking' perspective, as done for example, for major WMGHGs (den Elzen et al., 2005; Hohne et al., 2011) and NTCFs (Shindell et al., 2009), or to current



**Figure 8.1** | Cartoon comparing (a) instantaneous RF, (b) RF, which allows stratospheric temperature to adjust, (c) flux change when the surface temperature is fixed over the whole Earth (a method of calculating ERF), (d) the ERF calculated allowing atmospheric and land temperature to adjust while ocean conditions are fixed and (e) the equilibrium response to the climate forcing agent. The methodology for calculation of each type of forcing is also outlined.  $\Delta T_o$  represents the land temperature response, while  $\Delta T_s$  is the full surface temperature response. (Updated from Hansen et al., 2005.)

or projected future emissions in a ‘forward-looking’ view (see Section 8.7). Emissions estimates through time typically come from the scientific community, often making use of national reporting for recent decades.

With the greater use of emission-driven models, for example, in CMIP5, it is becoming more natural to estimate ERF resulting from emissions of a particular species rather than concentration-based forcing. Such calculations typically necessitate model simulations with chemical transport models or chemistry–climate models, however, and require careful consideration of which processes are included, especially when comparing results to concentration-based forcings. In particular, simulation of concentration responses to emissions changes requires incorporating models of the carbon cycle and atmospheric chemistry (gas and aerosol phases). The requisite expansion of the modelling realm for emissions-based forcing or emission metrics should in principle be consistent for all drivers. For example, as the response to aerosol or ozone precursor emissions includes atmospheric chemistry, the response to CO<sub>2</sub> emissions should as well. In addition, if the CO<sub>2</sub> concentration responses to CO<sub>2</sub> emissions include the impact of CO<sub>2</sub>-induced climate changes on carbon uptake, then the effect of climate changes caused by any other emission on carbon uptake should also be included. Similarly, if the effects of atmospheric CO<sub>2</sub> concentration change on carbon uptake are included, the effects of other atmospheric composition or deposition changes on carbon uptake should be included as well (see also Section 6.4.1). Comparable issues are present for other forcing agents. In practice, the modelling realm used in studies of forcing attributable to emissions has not always been consistent. Furthermore, climate feedbacks have sometimes been included in the calculation of forcing due to ozone or aerosol changes, as when concentrations from a historical transient climate simulation are imposed for an ERF calculation. In this chapter, we endeavour to clarify which processes

have been included in the various estimates of forcing attributed to emissions (Sections 8.3 and 8.7).

RF or ERF estimates based on either historical emissions or concentrations provide valuable insight into the relative and absolute contribution of various drivers to historical climate change. Scenarios of changing future emissions and land use are also developed based on various assumptions about socioeconomic trends and societal choices. The forcing resulting from such scenarios is used to understand the drivers of potential future climate changes (Sections 8.5.3 and 8.6). As with historical forcings, the actual impact on climate depends on both the temporal and spatial structure of the forcings and the rate of response of various portions of the climate system.

## 8.2 Atmospheric Chemistry

### 8.2.1 Introduction

Most radiatively active compounds in the Earth’s atmosphere are chemically active, meaning that atmospheric chemistry plays a large role in determining their burden and residence time. In the atmosphere, a gaseous chemically active compound can be affected by (1) interaction with other species (including aerosols and water) in its immediate vicinity and (2) interaction with solar radiation (photolysis). Physical processes (wet removal and dry deposition) act on some chemical compounds (gas or aerosols) to further define their residence time in the atmosphere. Atmospheric chemistry is characterized by many interactions and patterns of temporal or spatial variability, leading to significant nonlinearities (Kleinman et al., 2001) and a wide range of time scales of importance (Isaksen et al., 2009).