

## Frequently Asked Questions

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

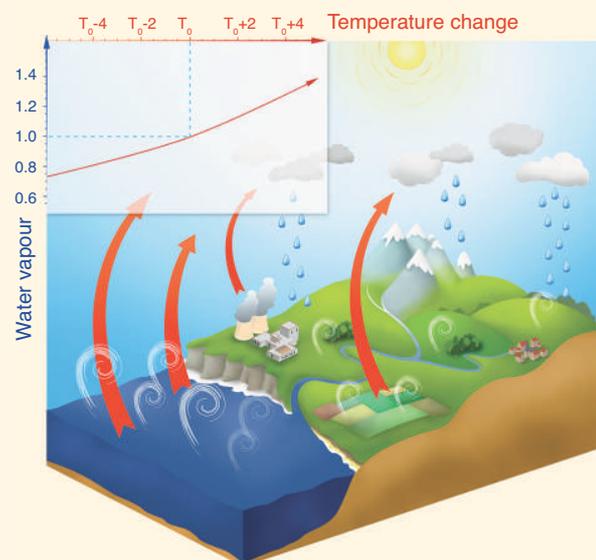
As the largest contributor to the natural greenhouse effect, water vapour plays an essential role in the Earth's climate. However, the amount of water vapour in the atmosphere is controlled mostly by air temperature, rather than by emissions. For that reason, scientists consider it a feedback agent, 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 forcers (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.