Water vapor and lapse rate feedbacks in the climate system

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Abstract

Water vapour is a greenhouse gas that dominates the Earth's terrestrial radiation absorption. As the planetary temperature warms, forced by increasing CO_2 and other greenhouse gases, water vapour content of the atmosphere increases, thereby producing the strongest positive feedback in the climate system. At the same time, the rate at which atmospheric temperature drops with height (the "lapse rate") is expected to decrease with warming. This represents a smaller, but significant, negative feedback, since it enables the planet to radiate more effectively to space. The two feedbacks are closely coupled to each other, and the "combined" result represents the foundational net positive feedback in the climate system, mandating substantial global warming in response to increased greenhouse gases.

This review summarizes the published work that has provided ever deepening understanding of these critical feedbacks. We outline the historical context, beginning with the 19th century awakening to the importance of water vapour in the climate, before focussing on the theoretical, observational and modelling work over recent decades that has transformed our understanding of their role in climate change. We show the evidence is now overwhelming that combined water vapour and lapse rate processes indeed provide the strongest positive feedback in the climate system. However important challenges remain. This review aims to provide physicists with a deeper understanding of these feedbacks, and stimulate engagement with the climate research community. Together the scientific community can provide further rigour, understanding and confidence in these most fundamental Earth system processes.

Contents

I. Introduction

Understanding and quantifying climate change is one of the outstanding scientific challenges of our era. The world is already seeing the impact of a changing climate, with approximately 1.0° C warming worldwide since pre-industrial times causing major impacts including loss of Arctic sea ice, rising sea levels, erosion of continental glaciers, melting permafrost, and increased incidence of extreme weather and climate including heatwaves, drought, fires, heavy rainfall and coastal flooding among many other changes (IPCC, 2013, 2014, 2019). Projections of warming over the rest of this century depend on future emissions of carbon dioxide, methane, nitrous oxide and other greenhouse gases (GHGs), but under a moderate emissions mitigation scenario are in the range 1-3° C, with a potential further warming of 1-2° C for a "business as usual" emissions (Collins et al., 2013).

Water vapour and lapse rate feedbacks play key roles in determining the magnitude of that warming. Water vapour feedback results from increasing moisture holding capacity of the atmosphere with temperature, diminishing the escape of outgoing terrestrial radiation. This dictates further warming to restore radiative equilibrium. At the same time, in latitudes spanning the tropics through to mid-latitudes, the upper troposphere warms faster than the surface – a change in the vertical "lapse rate" with temperature, enabling the Earth to radiate to space more effectively. This process offsets some, but most importantly not all, of the effects of increasing water vapour. Other feedbacks also operate involving changes in clouds, snow and sea ice, but the combined water vapour and lapse rate feedbacks can be considered to provide a fundamental amplification of climate warming, further enhancing the effects of other positive feedbacks.

Given their key importance for climate change, water vapour and lapse rate feedbacks have been the subject of intense research over the last four decades, and before that of study stretching back to the early 19th century. This research spans theoretical understanding of radiative impacts of humidity as climate warms, along with very extensive modelling and observational studies. This review paper will summarise this research and assess the current state of knowledge. It will also highlight areas where further rigour and understanding would provide even greater confidence in these critical Earth system processes.

The layout of this review is as follows.

Section II will provide a brief historical background of the understanding of the importance of water vapour in the climate system, and amplification of warming by water vapour feedback.

Section III will describe a formalism linking changes in "forcing" from GHGs with climate feedbacks and the response of the climate system.

Section IV will describe the radiative properties of water vapour that make it so important, and the fundamental distributions of temperature and water vapour in the atmosphere. It will describe the understanding of the manner in which spectral absorption by water vapour is related to changes in surface temperature, and what this implies for water vapour feedback. It also describes the unfolding of understanding of the importance of different regions in the atmosphere in setting the magnitude of both water vapour and lapse rate feedbacks. It will further discuss debates and research that have led to much deeper understanding of the processes controlling water vapour distribution in the current climate, and response in a warmer climate.

Section V lays out the observational evidence for strong positive water vapour, and negative lapse rate feedbacks, including evidence from climate variability, from climate change to date, from paleo climates, and from responses to volcanic eruptions.

Section VI will provide an assessment of Global Climate Model (GCM) representation of key physical processes, and a comparison with observed changes and variability to evaluate confidence in model water vapour and lapse rate feedbacks. Appendix 1 will also summarize methodologies for quantifying feedbacks, as these techniques have played an important part in the development of understanding and assessment.

Section VII gives perspective on the climate community understanding and consensus on these feedbacks, as expounded by evaluations carried out by the Intergovernmental Panel on Climate Change (IPCC) since the first report in 1990. The section will also describe and evaluate quantitative estimates of feedback strength, with details listed in Appendix 2.

Section VIII will provide a summary on the strength and consistency of evidence of the nature and magnitude of water vapour and lapse rate feedbacks. Finally, it will look to the future, to highlight remaining knowledge gaps and identify outstanding areas of further research.

II. Water vapour, lapse rate and the greenhouse effect.

A. A historical perspective

Understanding of the importance of the atmosphere for maintaining Earth's temperature through greenhouse trapping goes back nearly 200 years to Joseph Fourier, and his Insight that although the atmosphere is relatively transparent to incoming solar radiation, it strongly absorbs outgoing terrestrial radiation (Fourier, 1827). Laboratory measurements by John Tyndall (pictured in Fig. 1) later in the 19th century, established that the trace gases, water vapour and CO_2, were primarily responsible for the absorption of terrestrial radiation, rather than the primary atmospheric constituent gases of nitrogen and oxygen (Tyndall, 1861; Tyndall, 1872). Tyndall concluded that water vapour provided "a blanket, more necessary to the vegetable life of England than clothing is to man" (Fleming, 1998).

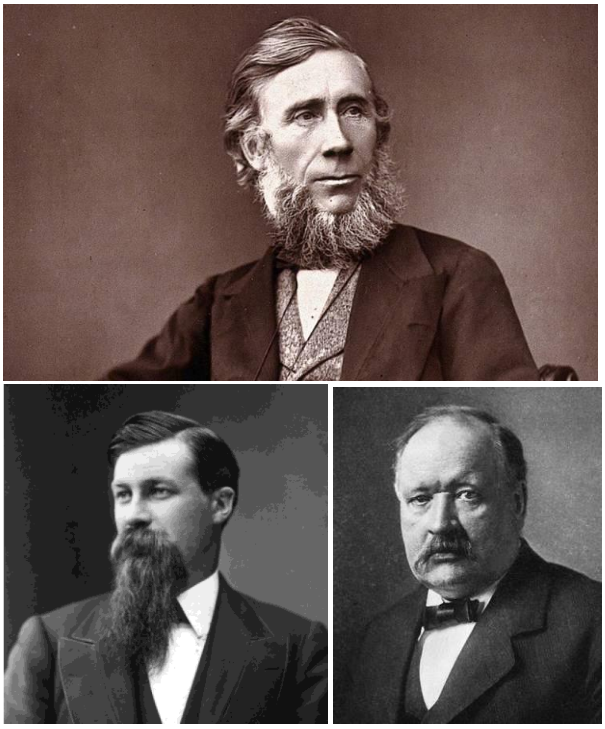


Figure 1: John Tyndall above, Thomas Chamberlin, below, Svante Arrhenius bottom right. Tyndall performed laboratory measurements of the absorption spectrum of water vapour in the mid-19th century. Thomas Chamberlin articulated the fundamental process controlling water vapour feedback. Arrhenius in the late 19th century laid out a coherent framework of CO_2 induced climate change, as amplified by water vapour feedback.

Svante Arrhenius (pictured in Fig. 1) wrote in 1896: "The selective absorption of the atmosphere … is not exerted by the chief mass of air, but in a high degree by aqueous water vapour and carbonic acid (CO_2) ...The influence of this absorption is comparatively small on the heat of the Sun, but must be of great importance in the transmission of rays from the Earth."

By the end of the 19th century there was an appreciation that water vapour could act as an amplifying "feedback" to other trace gas "forcing" (Arrhenius, 1896; Chamberlin, 1899). A statement essentially articulating the modern concept of water vapour feedback was made by Chamberlin (Fig. 1) in a letter to G.C. Abbott in 1905:

"[W]ater vapor, confessedly the greatest thermal absorbent in the atmosphere, is dependent on temperature for its amount, and if another agent, as CO_2, not so dependent, raises the temperature of the surface, it calls into function a certain amount of water vapor which further absorbs heat, raises the temperature and calls forth more vapor ..."

Twentieth century quantum theory has since provided theoretical understanding of the water vapour absorption spectrum, including the myriad of absorption lines due to rotational and vibrational absorption of infrared photons (see Section IV-A below). An additional "continuum" absorption, noted in the early 20th century (e.g. Brunt, 1932), is an important source of absorption between bands, but remains the least well understood component of the water vapour absorption spectrum (Shine et al. 2012), see Section IV-A.

Through the early to mid-20th century, further studies considered the quantitative role of water vapour feedback in determining response to CO_2 changes in the atmosphere (see Held and Soden, 2000, for an overview). Major advances occurred in the 1960s, with development of "one dimensional" models consisting of global mean profiles of temperature and moisture, with the temperature profile constrained not to exceed a specified "lapse rate", i.e. decrease in temperature with height. This lapse rate was considered to be set primarily by tropospheric convective processes (see Section IV-B). These so-called "radiative-convective models" (RCMs, Manabe and Strickler, 1964; Manabe and Wetherald, 1967) were able to provide a first-order representation of the troposphere[1], with a tropopause height determined by a combination of radiative and convective processes, and topped by a stratosphere in pure radiative equilibrium. These models were then used to estimate climate change induced by the addition of radiative absorbers such as anthropogenic CO_2 (Manabe and Wetherald, 1967; Ramanathan and Coakley, 1978).

A key advance in understanding the global response to CO_2 (and other GHG) increases, was the realisation that top of atmosphere (TOA) radiative balance, rather than surface radiation/evaporation balance dictated climate sensitivity (Manabe and Strickler, 1964; Manabe and Wetherald, 1967). An important additional step was the hypothesis that nearly unchanged relative humidity, rather than specific humidity, is more appropriate for climate change simulations (Manabe and Wetherald, 1967). Three-dimensional GCMs (latitude, longitude, height) replaced 1D models for climate change experiments from the 1970s, although the early RCMs shed crucial light on the role of water vapour feedback and lapse rate in climate sensitivity (see Kluft et al., 2019 for a recent analysis).

B. Radiative forcing and the "enhanced" greenhouse effect.

Let us consider what a very simplified RCM implies about global response to additional terrestrial wavelength absorbers (hereafter called "longwave", LW, absorbers). Incident TOA average annual solar radiation flux (vertical to the equatorial horizontal), S_0 is around 1,360 Wm^{-2} (Lacis et al., 2013). Since the atmosphere is largely transparent to incident solar radiation (see Section IV-A), most absorption takes place at the surface, modulated by the albedo (reflectivity) of clouds and Earth surface, including high albedo surfaces such as snow and sea ice. Observed global mean planetary albedo, α, is around 30% (Ramanathan, 2014). A straightforward calculation gives net absorbed solar (hereafter called "shortwave", SW) radiation of $\frac{S_0}{4}(1-\alpha)$ or approximately 240 Wm^{-2}.

[1] The troposphere is the bottom roughly 6-10km of the atmosphere, and of generally decreasing temperature with height. Weather events are confined to this zone. It is separated from the overlying stratosphere by the tropopause.

Most LW radiation from the surface cannot escape directly to space due to the opacity of the atmosphere (see Section IV-A), and multiple absorptions and emissions culminate in an effective radiating height, Z_e, of around 500 hPa (or approximately 5 km) at a global mean effective radiating temperature, T_e, of around 255K. Assuming a fixed lapse rate Γ, corresponding to the observed global mean value of ~6.5K/km, gives a surface temperature of $T_s = T_e + \Gamma Z_e$, or approximately 288K in the pre-industrial era. This simple calculation describes a "natural greenhouse effect" of the atmosphere, resulting in a planetary surface temperature that is around 33K warmer than that with no atmosphere (Lacis et al., 2013).

Adding infrared absorbers, such as CO_2, CH_4 or N_2O increases the atmosphere's opacity, forcing an increase in Z_e (Fig. 2). Assuming an unchanged lapse rate, a doubling of CO_2 raises Z_e by approximately 150m (Held and Soden, 2000) and thus increases the surface temperature by 150m x 6.5K/km, or ~1K. This warming, uniformly spread throughout the atmosphere and assuming a black body radiation according to the Stefan-Boltzmann law, increases the upward radiative flux at the TOA by ~4Wm^{-2}, thereby balancing the reduced TOA outward flux induced by the increased CO_2.

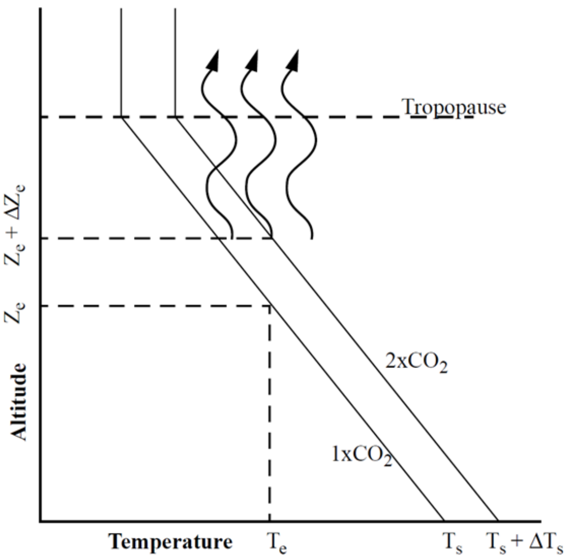


Figure 2: A schematic illustrating how additions in atmospheric greenhouse gases, such as a doubling of CO_2 concentration, changes surface temperature. Assuming an unchanged lapse rate, additional longwave absorbers and unchanged emission temperature, T_e, forces radiation to space to come from a higher altitude, increasing surface temperature, T_s. Source: Held and Soden (2000).

This simple paradigm can be considered a "no feedback response", with only the vertically uniform "Planck" radiative damping operating (Bony et al, 2006). This provides a first order understanding of the planetary response to increased GHGs, with feedbacks including water vapour, lapse rate, clouds and surface albedo then operating in addition to this basic response. These further modify Z_e, Γ, T_e and T_s, through changes in the absorption/reflection of downward solar radiation, in lapse rate and in the strength and vertical distribution of additional LW absorption (Held and Soden, 2000). Understanding these feedbacks, their underlying physical processes, their magnitude and their interactions have been among the principal goals of climate research over the last five decades as these set the fundamental sensitivity to greenhouse gases (e.g. Bony et al., 2015).

As noted above, early studies recognized the potential significance of the strong temperature dependence of the equilibrium vapour pressure of water as a feedback mechanism, but lacked a clear quantification of its importance. A key insight from the RCMs was that if relative humidity stayed close to constant, water vapour feedback roughly doubled the "no feedback" warming described above (assuming no change in the lapse rate). The landmark study of Manabe and Wetherald, (1967) deduced a consequent global surface temperature response to CO_2 doubling of 2.3K, a value well within the range of modern GCMs, and had profound influence on subsequent research[2]. Of course, this was a 1D model only, ignoring many processes, such as the general circulation of the atmosphere, differing tropical/extra tropical

[2] A 2015 survey of climate scientists voted it as the most influential climate change paper of all time, being the "first proper computation of global warming … from enhanced greenhouse gas concentrations, including atmospheric emission and water-vapour

regions, ocean circulation, snow, sea ice, and land surface processes. Indeed, the assumption of constant relative humidity was born more out of necessity than strong theoretical or empirical support. Subsequent studies with 3D models would show that relative humidity does exhibit systematic changes regionally (Sherwood et al., 2010a,b), however at the global scale, the strong temperature dependence overwhelms the influence of regional variations in relative humidity. The basic tenets of water vapour feedback strength from 3D models substantiate the early 1D model estimates (Colman, 2001; Soden and Held, 2006; Boucher et al., 2013). Given this central importance in amplifying anthropogenic climate change, water vapour feedback, along with the associated lapse rate feedback has undergone intense scrutiny over the last three decades from theoretical, observational, modelling and process studies.

III. Global radiative feedbacks and climate sensitivity.

A. A global feedback paradigm

The longstanding paradigm within the climate community for understanding the equilibrium climate response to forcing (e.g., Hansen, 1984; Sherwood et al., 2015) has been adapted from the classic model of the response of an electronic amplifier to perturbation (Bode, 1945).

Consider a radiative perturbation or "forcing" to the climate system, such as from a change in atmospheric CO_2, that instantaneously affects the TOA[3] radiative balance by an amount ΔF. Under the radiative imbalance, T_s responds, and with it a myriad of possible processes may be affected in the atmosphere and at the planetary surface which in turn affect TOA radiation balance, R, either by changing the outgoing LW radiation (OLR) or the SW reflected radiation. Assuming that *the net effect of the processes is related to global mean surface temperature*[4], we can then write:

$$\Delta R = \Delta F + \lambda \Delta T_s \qquad (1)$$

where λ is defined as the *climate feedback parameter* and has units of $Wm^{-2}K^{-1}$. Here we define the radiative flux[5] as downwards positive (i.e. warming), although in fact there is no universal convention in the climate literature. Taking x as a vector of processes affecting R, following the formulation of Bony et al. (2006) and Knutti and Rugenstein (2015) we formally define the feedback parameter as

$$\lambda = \frac{\partial R}{\partial T_s} = \sum_x \frac{\partial R}{\partial x}\frac{\partial x}{\partial T_s} + \sum_x \sum_y \frac{\partial^2 R}{\partial x \partial y} + \ldots \qquad (2)$$

In the traditional feedback formulation, the most "fundamental" response of the climate system, analogous to open-loop gain in the electronic context, is the so-called "Planck response", $\Delta T_{s,P}$. This is where the surface temperature and the overlying atmospheric temperature respond uniformly with height – *with all other atmospheric and surface variables unchanged* (e.g. Bony et al., 2006). Assuming the Planck response only,

$$\Delta R = \Delta F + \Delta T_{s,P}(\lambda_P), \qquad (3)$$

where $\lambda_P = \frac{\partial R}{\partial T_{s,P}} \approx -4\sigma T_e^3 \approx -3.2 Wm^{-2}K^{-1}$

where σ is the Stefan–Boltzmann constant. Note that this value of λ_P, obtained by simply differentiating the Stefan-Boltzmann law is notably smaller than that calculated by climate models, ~4 $Wm^{-2}K^{-1}$, primarily due to the lack of stratospheric warming due to its decoupling from the surface (Cronin 2020). A doubling of atmospheric CO_2, corresponding to a radiative forcing of

feedback" https://www.carbonbrief.org/the-most-influential-climate-change-papers-of-all-time.

[3] The imbalance is often specified at the tropopause rather than the TOA, but the difference between formulations is trivial since the stratosphere equilibrates to forcing on timescales of a few weeks – essentially instantaneous for climate change considerations – thereby equalising TOA and tropopause imbalances (Hansen et al., 1984).

[4] Other assumptions are possible – a recent proposal that feedbacks be better related to mean tropical 500hPa temperatures (Dessler et al, 2018) provides a different feedback formulation. This approach will not be discussed further however as little literature is yet available on feedbacks under this formalism.

[5] In the climate literature, the radiative *flux density* is typically referred to simply as radiative *flux*. In this paper, we use the term *radiative flux* to refer to the spectrally integrated radiative power per unit area in units of Wm^{-2}.

approximately 4 Wm^{-2}, produces surface warming of around 1.2K (Bony et al., 2006). Note that although *horizontal* uniformity of $\Delta T_{s,P}$ is often also assumed for the Planck response, little difference occurs to this calculation if the temperature response varies geographically. For example, no fundamental difference occurs if the "Planck" warming is enhanced at high latitudes, as is normally the case in Global Climate Model (GCM) warming response to CO_2 forcing (Colman, 2004; Section IV-I). Alternative "no-feedback" vertical temperature profiles have been proposed, other than a uniform increase with height (e.g. Schlesinger et al., 2012) but have not come into common usage, so will not be discussed further here.

In the presence of non-Planck processes, such as temperature dependent changes to water vapour, lapse rate or clouds we can express final surface temperature change as $\Delta T_s = \left(\frac{\lambda_P}{\lambda}\right)\Delta T_{s,P}$. Ignoring second and higher order terms in Eqn 2, we write

$$\lambda = \lambda_P + \sum_{x \neq P} \lambda_x \ , \qquad (4)$$

where $\lambda_x = \frac{\partial R}{\partial x}\frac{\partial x}{\partial T_s}$ and where $x \in \{q, \Gamma, \alpha, C\}$,

corresponding to water vapour, lapse rate, surface albedo and cloud feedbacks respectively. These are commonly referred to as the "fast" feedbacks of the climate system, as they respond to surface temperature changes on rapid timescales, in the case of water vapour, lapse rate and clouds, on the order of hours to weeks, much shorter than, for example, adjustment timescales of the ocean. Beyond the fast feedbacks lie many other processes that (eventually) impact radiation. These include land and ocean carbon cycle feedbacks, ecosystem responses, vegetation albedo feedbacks and many others which affect GHG concentration and TOA radiative balance (see Heinze, et al., 2019; Tierney et al., 2020 and references therein). These are important for long (multi-decadal or longer) timescale earth system responses, and will interact with fast feedbacks (Heinze, et al., 2019), but are beyond the scope of this review.

For water vapour, given the close to logarithmic dependence of LW radiation on specific humidity, and the roughly exponential rate of increase of saturation specific humidity with temperature (see section IV-B) a pragmatic alternative can be to instead use $\lambda_q = \frac{\partial R}{\partial \ln q}\frac{\partial \ln q}{\partial T_s}$ (Raval and Ramanathan, 1989; Soden and Held, 2006). This formulation has been widely used in the calculation and application of radiative "kernels" used for evaluating feedbacks in practise (see Appendix 1).

Normalising each feedback by the Planck response defines the "gain" from each feedback, $g_x = -\frac{\lambda_x}{\lambda_P}$, allowing us to write:

$$\Delta T_s = \frac{\Delta T_{s,P}}{1 - \sum_{x \neq P} g_x} \ . \qquad (5)$$

Positive feedbacks, then, are viewed as those that "oppose" the Planck cooling, reducing the effectiveness of the planet below its simple blackbody cooling rate, and therefore amplifying the global mean surface temperature response required to re-achieve TOA balance. When ΔF corresponds to a doubling of atmospheric CO_2 concentrations, the equilibrated temperature $\Delta T_{2\times CO2} \equiv -\frac{\Delta F_{2\times CO2}}{\lambda}$ is referred to as the "Equilibrium Climate Sensitivity" (ECS). ECS therefore considers only so-called "fast" climate feedbacks. ECS, although an idealised concept, never fully realised in the real world, has been talismanic in climate science for more than 40 years as a benchmark measure of climate change (Charney et al., 1979; Sherwood et al., 2020) and remains an extremely useful parameter, for example being proportional to the transient rate of warming projected by GCMs over the 21st century (Gregory et al., 2015; Grose et al., 2018).

Observational and modelling studies, discussed in this review, find that the water vapour and lapse rate feedbacks amplify global warming from CO_2 and other GHGs by a factor of ~2, with a total gain of ~0.5 (Bony et al., 2006; Randall et al., 2007). Other feedbacks in the climate system, due to clouds and surface albedo (predominantly snow and sea ice) then operate "on top of" this enhanced warming and amplify or damp the response further. The critical nature of the combined water vapour+lapse rate feedback is apparent from this formalism. Due to the non-linearity in Eqn 5, the 0.5 gain from water vapour and lapse rate acts to "sensitise" the climate response, giving greatly boosted warming from further positive feedback such as due to surface albedo or clouds (Bony et al., 2006; Zelinka et al., 2020).

The nonlinear nature of feedback contribution to climate sensitivity in Eqn 5, means that apportioning fractional climate change (in this case of global mean surface temperature change) to individual feedbacks, depends on the state of all other feedbacks (Held and Soden, 2000; DuFresne and Bony, 2008). Similarly, the *uncertainty* caused by any one feedback has the same state dependency. Therefore, although it is acknowledged that there is no unique way to achieve this subdivision, a useful methodology proposed by DuFresne and Bony (2008) follows the above "gain" approach of normalising each feedback by the strength of the Planck response, thereby reiterating the "differing nature" of the Planck response compared to the other feedbacks. Figure 3 shows the results for 12 models of the Coupled Model Intercomparison Project, phase 3 (CMIP3, Meehl et al., 2007) ensemble. It shows a measure of the relative importance of the combined water vapour+lapse rate feedback on global temperature change and its uncertainty. Clouds dominate overall projection uncertainty, although water vapour+lapse rate remains the second most important contributor, and provides the greatest addition to temperature change over the basic Planck response.

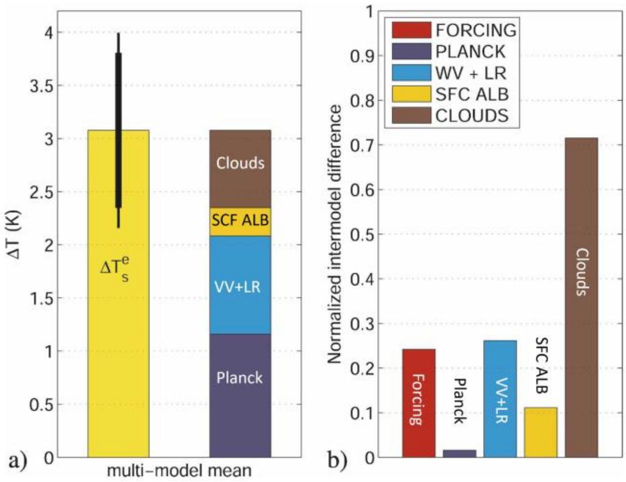


Figure 3. (Colour online) (a) CMIP3 multi-model mean surface temperature change ΔT^e_s (equivalent to ΔT_s in Eqn 5) under a doubling of CO_2, thick and thin lines represent 1 and 2 standard deviation range. Coloured bars show multi-model mean contribution to ΔT^e_s from each of the feedbacks listed, according to "gain" factors in Eqn 5. (b) Contribution to range in ΔT^e_s of the different feedbacks, calculated as the standard deviation of contribution to temperature change, normalised by ΔT^e_s. Source: DuFresne and Bony (2008). © American Meteorological Society. Used with permission.

B. Alternate feedback formulations

Traditionally, water vapour and lapse rate feedbacks were considered as separate processes (e.g. Schlesinger, 1988), however they are closely linked, and much can be gained from considering them as a combined feedback (Soden and Held, 2006; Held and Shell, 2012; Ingram 2010, Po-Chedley et al., 2018). As discussed in Section IV-E, a strong anticorrelation is found between water vapour and lapse rate feedbacks. Two approaches have been adopted to take consideration of this strong anti-correlation. The first is to simply sum the two feedbacks, resulting in a combined water vapour+lapse rate feedback – e.g. as assessed by the Intergovernmental Panel on Climate Change (IPCC) in the 5th Assessment Report (Boucher et al., 2013) and discussed above when considering the relative contributions of feedbacks to final temperature change (Fig. 3).

An alternative formulation from Held and Shell (2012), and drawing from earlier work by Simpson (1929) and Ingram (2010, 2013a), posits that the assumption of a Planck response with unchanged specific humidity is fundamentally unphysical. This is because it implies large relative humidity drops with increasing temperatures, which are not seen either in observations (see Section V) or in GCMs (see Section VI). The subsequent "restoration" of an unchanged relative humidity with the Planck warming, which forms a large part of the water vapour feedback, is then seen as a physically artificial adjustment, leading by construct to a strong positive water vapour feedback, in turn opposed by a strong negative lapse rate feedback. Instead, a more "fundamental" Planck response can be considered one of fixed relative humidity. Under this assumption, following Held and Shell (2012), we construct a modified Planck feedback:

$$\lambda'_P = \lambda_P + \lambda_{Pq} \quad (6)$$

where λ_{Pq} corresponds to the radiative response from the water vapour changes required to maintain fixed relative humidity under the (vertically uniform) Planck temperature response. Under this formulation, the lapse rate feedback now also includes the radiative response from ensuring fixed relative humidity under the changed lapse rate, viz: as:

$$\lambda'_\Gamma = \lambda_\Gamma + \lambda_{\Gamma q} . (7)$$

The reformulated "water vapour feedback" now includes only relative humidity changes:

$$\lambda'_H = \lambda_q - \lambda_{Pq} - \lambda_{\Gamma q} . \quad (8)$$

Surface albedo and cloud feedbacks are unaffected by this transformation. The surface temperature response is now expressed as:

$$\Delta T_s = \frac{\Delta T_{s,P'}}{1 - \sum_{x \neq P} g'_x} , \qquad (9)$$

where, $g'_x = -\frac{\lambda'_x}{\lambda'_P}$ and $x \in \{H, \Gamma, \alpha, C\}$, where the symbols are as for Eqn 4, with H representing the relative humidity and $\Delta T_{s,P'} = -\frac{F}{\lambda'_P}$ (Held and Shell, 2012).

A comparison of traditionally defined Planck, water vapour and lapse rate feedbacks, and the relative humidity transformed feedbacks for the CMIP3 model ensemble are shown in Fig. 4 (Boucher et al., 2013). It is immediately apparent that the relative humidity formulation removes the large offsetting feedbacks, and reduces the inter-model spread.

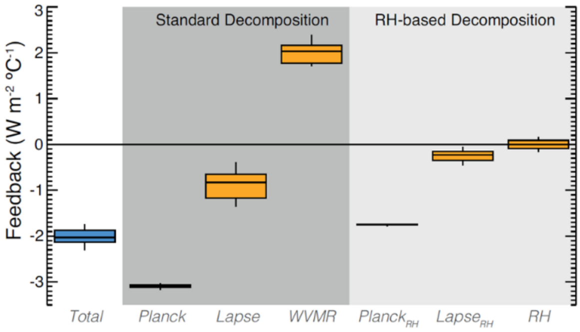


Figure 4. (Colour online) Feedback parameters associated with water vapour or the lapse rate predicted by CMIP3 GCMs, with boxes showing interquartile range and whiskers showing extreme values. At left is the total radiative response including the Planck response. In the darker shaded region is the traditional breakdown of this into a Planck response and individual feedbacks from water vapour (labelled "WVMR") and lapse rate (labelled "Lapse"). In the lighter-shaded region at right are the equivalent three parameters calculated in the alternative, RH-based framework. In this framework all three components are both weaker and more consistent among the models. (Data are from Held and Shell, 2012). Source: Boucher et al. (2013).

In this review, we contend that both the traditional formulation, and the relative humidity-based formulation (hereafter called "RH feedbacks") are useful, providing different insights into the nature of feedbacks, their importance in determining large-scale response to forcing and the nature and importance of feedback spread (e.g. Ingram, 2013b). Furthermore, within the traditional framework, the consideration of separate water vapour and lapse rate feedback, and their simple sum are both useful approaches in different contexts. If the sum of water vapour and lapse rate feedback agrees between models, then this provides a pragmatic approach to narrowing uncertainty in ECS and focuses research concentration on cloud and surface albedo feedbacks which have greater impact on ECS uncertainty. However, if different models produce the same net feedback balance through different mechanisms, this undermines confidence in models generally, and specifically for aspects of projections dependent on model representation of those processes. In practical terms too, the overwhelming majority of studies over the last 40 years have considered traditional, separate feedback, so the focus of the research community has been heavily fixed on the traditional definitions. Finally, the sum of the modified lapse rate feedback (Eqn 7), the RH feedback (Eqn 8) and the term due to the humidity increase under vertically uniform temperature increase with fixed relative humidity (λ_{Pq} in Eqn 6), equals the traditionally defined water vapour+lapse rate feedbacks.

It may be asked: how "separable" are feedbacks in practise i.e., can they be divided into separate processes in practise, not just theory? Strong support for this is provided by GCM studies where individual feedback loops are "cut" by suppressing their radiative impact (see Appendix 1). Mauritsen et al. (2013) found for one such GCM that this yielded a "near-perfect decomposition of change into partial temperature contributions pertaining to forcing and each of the feedbacks", including a separation of water vaper and lapse rate feedbacks according to the traditional framework.

C. Feedbacks at the Earth's surface

The discussion to date has focused on TOA forcing and feedback, as TOA radiative balance is fundamental to planetary equilibrium and response to forcing (Manabe and Wetherald, 1967).

It can also be useful, however, to consider feedbacks from a surface perspective which can provide additional insights into radiative impacts on processes such as evaporation, with consequences for changes to atmospheric temperature and rainfall (Andrews et al., 2009; Previdi, 2010). Under a small climate perturbation, the surface net radiative budget can be written as:

$$\delta R = \delta R_P + \delta R_q + \delta R_\Gamma + \delta R_C + \delta R_{RF} \qquad (10)$$

where R is now the net surface radiation, RF the surface radiative forcing, and the other surface radiation terms are from changes in the Planck term, water vapour, lapse rate, surface albedo and clouds. Ignoring the small heat conduction term into the soil, net surface heat balance, W, can be written

$$\delta W = \delta R + \delta E + \delta S, \qquad (11)$$

where E and S represent the evaporative and sensible heat turbulent fluxes respectively (Colman, 2015). Surface feedbacks are discussed in Section IV-H below.

IV. Physical processes

A. Radiative properties of water vapour

Water vapour has a profound impact on Earth's outgoing LW radiation. It is responsible for around 50% of total absorption and around 60% of the total clear sky[6] "greenhouse effect" in the near infrared (Kiehl and Trenberth, 1997; see Fig. 5). Being a strongly polar molecule (in contrast to CO_2 for example) it has numerous absorption modes from rotation in three separate axes. These rotational modes combine with vibrational modes, producing a huge number of absorbing bands in the near and mid-infrared. Molecular bending and symmetric/asymmetric stretching contribute to other absorption modes, often overlapping with tones and overtones of other modes (Stevens and Bony, 2013). The result is "bands" consisting of thousands of closely packed narrow absorption lines (Goody and Robinson, 1951; Fig. 5). In addition, throughout the spectrum from the microwave through to the visible, lies a relatively smoothly varying absorption "continuum" (Brunt, 1932; Clough et al., 1989, Tipping and Ma, 1995). This continuum is particularly important in the "window" zones between the bands, where it is the dominant source of absorption (Shine et al., 2012; Stevens and Bony, 2013; Lechevallier et al., 2018).

[6] i.e., radiation calculations performed in which cloud amounts are set to zero.

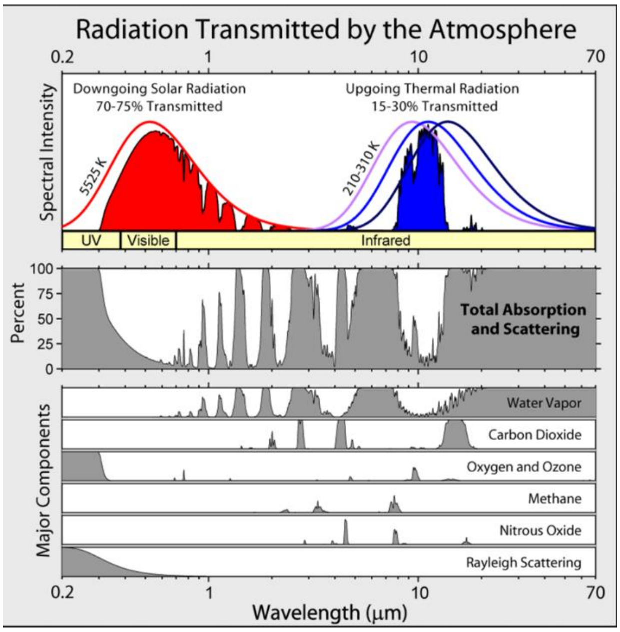


Figure 5. (Colour online) Absorption spectra for total atmosphere and water vapour, CO_2 and other atmospheric gases, as a function of wavelength. Also shown are the "blackbody" curves of downward solar (SW) radiation and upward terrestrial (LW) radiation. Source: https://upload.wikimedia.org/wikipedia/commons/7/7c/Atmospheric_Transmission.png and following Goody and Robinson (1951).

The source of the continuum has been debated for several decades, see review by Shine et al. (2012) and references therein, with candidate mechanisms including "far-wing" effects from remote spectral lines and absorption by dimers. Uncertainties in the details of the physics underlying the continuum, albeit still an important research topic in the molecular spectroscopy community, have little impact on the strength of water vapour feedback in the infrared (Huang et al. 2007). This is of course so long as GCMs parameterise the essential features of both band and continuum absorption, as well as radiation codes in support of observations, including satellite retrieval of features such as upper tropospheric humidity (Soden et al., 2000). To this end, model radiation codes have been compared with very detailed and sophisticated "line by line" radiation calculations in several major inter-model comparisons (Pincus et al., 2016). It is found that limitations in representation of radiation do not represent a material uncertainty in TOA radiation balance or water vapour feedbacks in models (Allan, 2012).

Typically, to a reasonable approximation, the "saturation" of large parts of the water vapour spectrum mean that addition of extra water vapour does not increase absorption in the central part of absorption lines, but only in their far "wings". This means that absorption is proportional to the logarithm of specific humidity (Lacis et al., 2013). This is a key point that implies that absorption depends on *relative* humidity changes as temperatures increase (see below).[7]

In the SW, although the (cloud free) atmosphere is mostly effectively transparent, water vapour is also the most important atmospheric constituent, being responsible for more than 60% of the total absorption by atmospheric gases (Kiehl and Trenberth, 1997; Fig. 5). There remain some questions in GCM representation of the SW absorption by water vapour. These uncertainties, particularly in the near-infrared, may have important implications for the atmospheric energy balance and thus how precipitation changes in response to a moistening climate (Radel et al., 2015; DeAngelis et al., 2015). The uncertainties are less important for water vapour feedback (Takahashi, 2009; Allan et al, 2012), although there may be links to climate sensitivity through cloud impacts (Watanabe et al., 2018).

In summary, then, the radiative properties of water vapour are well understood, and their representation in models is sufficiently accurate as to rule out their contributing any significant uncertainty to water vapour feedback.

B. Fundamentals of temperature and water vapour distributions in the atmosphere

1. Lapse rate

With limited exceptions, temperature decreases with height in the troposphere, due to the absorption

[7] Note that there are some modest departures to the logarithmic absorption dependence: remaining unsaturated weak lines have a linear dependence on specific humidity (Lacis et al., 2013), and continuum absorption in the spectral "windows" increases as the square of water vapour density (Baranov et al., 2008).

of the majority of solar radiation at the surface and atmospheric cooling with altitude from expansion as parcels rise. Parcels raised from the surface, with no input/loss of heat (termed an adiabatic process), cool at the "adiabatic lapse rate", for dry air given by $\Gamma_d = g/c_p \approx 10Kkm^{-1}$, where g is the gravitational constant and c_p the specific heat capacity of air. Another way of thinking of this is the conservation of "dry static energy", $S = c_p T + gz$, where z is height above the surface.

The presence of water vapour, however, changes this profoundly. As water vapour in the parcel reaches saturation, further ascent results in condensation, with latent heat release offsetting some of the cooling. The moist or "saturated" lapse rate is given by:

$$\Gamma_m \approx \Gamma_d \left[\frac{1 + \frac{e^*}{P}\left(\frac{\beta}{T}\right)}{1 + \frac{e^*}{P}\frac{R_d}{c_p}\left(\frac{\beta}{T}\right)^2} \right] \quad (12)$$

where e^* is the saturation vapour pressure (see Eqn 13 below), P the pressure, T temperature, R_d the dry air specific gas constant and β a factor roughly equal to the ratio of latent heat of vaporisation at constant pressure to the water vapour gas constant (Stevens and Bony, 2013). Near the surface $\Gamma_m \approx 4Kkm^{-1}$ but approaches Γ_d aloft, where (saturation) vapour pressure becomes very low as a result of very low temperatures. In moist environments, if atmospheric lapse rates exceed saturated adiabatic lapse rate, convection will act to stabilise the atmosphere. Above the tropopause (the upper limit of convection), the stratosphere lies in close to true radiative equilibrium, and shows largely unchanged or slightly increased temperature with height.

In the tropics, the atmosphere is observed to be close to saturated adiabatic across broad regions (Xu and Emanuel, 1989; Sobel et al., 2001). This is due to convective stabilisation in moist regions along with the fact that the Coriolis effect is small here. The latter means that dynamical circulations quickly erode horizontal temperature gradients, so the lapse rate broadly is set by the areas with the deepest convection (Neelin and Held, 1987; Lambert and Taylor, 2014).

In mid-latitudes baroclinic adjustment (associated with extra-tropical cyclones, anticyclones and planetary scale waves) is a key process setting lapse rate (Stone, 1978; Stone and Carlson, 1979), although with some seasonal variation. In the summer hemisphere, in particular, convective cores within the warm parts of baroclinic eddies can result in a lapse rate similar to moist adiabatic (Juckes, 2000), with implications for amplified upper tropospheric warming under global temperature increase (Frierson, 2006).

At high latitudes the most common conceptual model of the basic state is one of "radiative-advective" equilibrium, with a balance between heat flux convergence from lower latitudes (by atmospheric and or oceanic processes), balanced by absorbed SW radiation, and LW cooling (Payne et al. 2015; Cronin and Jansen, 2016). Vertical temperature profiles are set by the balance between surface and atmospheric SW absorption and commonly result in temperature inversions and stable atmospheric profiles – a cold surface layer overtopped by warmer air. The role of lapse rate feedback in controlling high latitude warming under radiative forcing is discussed in detail in Section IV-I.

2. Water vapour

Despite its profound radiative impact, water vapour accounts for only around 0.25% of atmospheric mass (Stevens and Bony, 2013). For perspective, if all water in the atmospheric column were precipitated, it would represent a globally averaged depth of only around 2.5 cm, dwarfed by an oceanic depth (globally distributed) of around 2.8 km. The vapour (gas) state comprises more than 99% of total atmospheric water content (Stevens and Bony, 2013).

In a given air parcel, water vapour pressure in the presence of liquid water (such as water droplets) represents a balance between departure of individual molecules from the water surface and collision and coalescence of molecules within the surface. The departure process, in particular, is highly temperature dependent. When these rates are matched, the atmosphere is saturated with respect to water vapour. This equilibrium or saturation vapour pressure e^* is described by the Clausius-Clapeyron equation:

$$\frac{\mathrm{d}e^*}{\mathrm{d}T} = \frac{Le^*}{R_v T^2} \qquad (13)$$

where T is a temperature, R_v is the gas constant for water and L is the latent heat of vaporisation (or sublimation). Specific humidity, h, used below, is related to the vapour pressure by $h = {}^{e}\!/_{R_v T}$. Relative humidity is given by e/e^*. Equation 13 implies that rate of increase of saturation specific humidity with temperature is itself a function of temperature, increasing from 6% per K at 35° C, to 7% per K at 0° C, and to 17% per K with respect to ice at -85° C (around the coolest tropospheric temperatures, occurring near the tropical tropopause). This represents roughly a doubling with every 10° C (Lacis et al., 2013). Note that super saturation with respect to liquid water is rare in the atmosphere due to an abundance of condensation nuclei (Sherwood et al., 2010b), although such is not the case with ice saturation (Jensen et al., 2005). Given the overall strong decrease of temperature with height, specific humidity drops by more than four orders of magnitude from the (tropical) surface to the tropopause.

3. Relative humidity distribution

Despite *specific humidity* decreasing roughly exponentially with height, *relative humidity* follows a very different profile, with large values in the boundary layer, reducing to minima in the subtropical mid-troposphere, then increasing again above that (Fig. 6). In the deep tropics relative humidity is high throughout the troposphere with a secondary maximum around 200-300hPa. In mid to high latitudes relative humidity decreases with height, but only slowly, with high values persisting well into the mid-troposphere (Fig. 6).

An analytical model of the tropical troposphere by Romps (2014) was able to describe the main features of the vertical humidity profile. In the lower troposphere, decreasing relative humidity with height resulted from decreasing convective detrainment[8] coupled with subsidence "drying" (i.e. decrease in relative humidity as air parcels warm and specific humidity is unchanged). On the other hand, the increase of relative humidity with height towards the upper troposphere was from increased fractional convective detrainment which increases rapidly as the mass flux (i.e. total upward air transport) of convective systems dwindles (Romps, 2014). Mid latitude mixing also plays an important role (Galewsky et al., 2005), and these processes will be described in more detail in the next section. Climate models overall can reproduce the features of large-scale relative humidity distribution with significant skill (Fig. 6; Bates and Jackson, 1997; Gaffen et al., 1997; Randall et al., 2008; Flato et al., 2013). Overall, too, climate models show skill in reproducing observed mean lapse rate in the tropics and elsewhere (Flato et al., 2013).

The principal questions that follow now are how do the distributions of temperature and water vapour change in a warming climate, say initiated by increases in CO_2 or other GHGs, and how do these affect the TOA radiation?

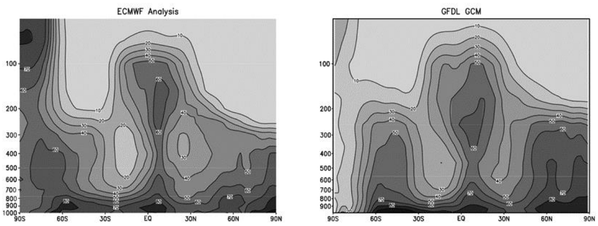


Figure 6. Zonal mean distributions of relative humidity as a function of latitude and height in the European Centre for Medium-range Weather Forecasts (ECMWF) reanalysis for July 1987 (left) and in a single GCM (right), that of the Geophysical Fluids Dynamics Laboratory. Vertical unit is hPa. Source: Held and Soden (2000).

C. Spectrally-Dependent Response to Warming

Early studies of the outgoing terrestrial radiation (Simpson, 1929) noted a surprising insensitivity of the spectrally-integrated outgoing longwave radiation to surface warming if the atmosphere was allowed to moisten while conserving both lapse-rate and relative humidity, sometimes referred to as *Simpson's Paradox* (Jeevanjee, 2018). Such

[8] "Detrainment" refers to the mixing of (often saturated) air within convective towers into surrounding air, causing increases in humidity in the region around the convection.

insensitivity is both counterintuitive, given the Planck functions strong dependence on temperature, and physically unrealistic, as it places the climate system in a perpetual runaway configuration (Nakajima et al., 1992; Pierrehumbert, 2010).

Resolution of this paradox comes by accounting for the spectral dependence of water vapour absorption, information that was not available at Simpson's time, as well as the influence of other absorbers, such as CO_2 and clouds, on the atmospheric emission (Ingram, 2010). Indeed, further insight into the water vapour and lapse-rate feedbacks can be gained by separating the outgoing longwave emission into two regions: one where water vapour absorption is optically thick and outgoing emission is insensitive to surface warming when relative humidity is conserved, and the other where it is optically thin and emission from the surface and atmosphere closely follows that of a blackbody (Ingram, 2013a; Jeevanjee, 2018), sometimes referred to as a partly-Simpsonian model.

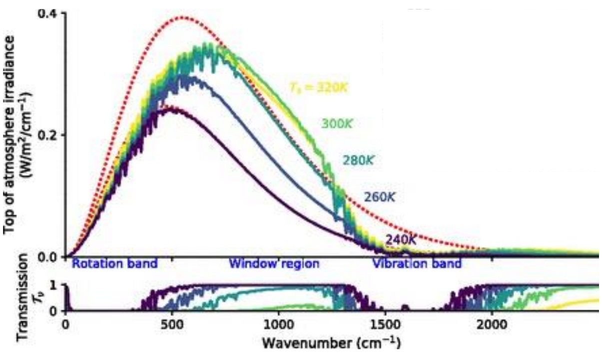


Figure 7. (Top) Calclulations of the spectrally-resolved OLR as a function of temperature for idealized atmospheric profiles in radiative convective equilibrium with constant relative humidity (r=100%). Red curves show the surface's blackbody emission at 240 K and 280 K. (Bottom) Spectrally resolved transmission between the surface and top of atmosphere for each profile. Source: Koll and Cronin (2018). Reproduced with permission PNAS.

Figure 7, from Koll and Cronin (2018), shows the spectrally-resolved outgoing terrestrial radiation for a set of atmospheric profiles with constant relative humidity and moist adiabatic lapse rates, but varying surface temperature. Note that in the optically thick regions of the water vapor rotational (wavenumber 1/l < 500 cm^{-1}) and vibrational (1/l > 1500 cm^{-1}) absorption bands, emission changes little with surface warming and the atmospheric transmissivity is near zero. Jeevanjee et al. (2021) show that this insensitivity arises from a near-perfect cancellation between changes in black body emission and attenuation by water vapour at these wavelengths, validating much of Simpson's original premise.

The large compensation between lapse-rate and water vapour feedbacks originates from the tendency for models to conserve relative humidity, resulting in this spectral cancellation between emission and attenuation within the water vapour absorption bands. As noted by Jeevanjee et al. (2021), this cancellation is eliminated if the water vapour and lapse-rate feedbacks are reformulated into the alternative relative humidity-based framework (*cf*. Fig. 4), enabling greater insight into the processes responsible for the differences in climate sensitivity between models (Po-Chedley et al., 2018; Zelinka et al., 2020; Zhang et al., 2020a; He et al., 2021).

There are two basic consequences of the partly-Simpsonian behavior of the atmosphere. The first is that the stabilization of Earth's climate to surface temperature change is achieved almost exclusively through radiative damping within the atmospheric window where water vapor absorption is negligible (Slingo and Webb, 1997; Koll and Cronin, 2018; Seeley and Jeevanjee, 2021).

This can be illustrated using the partly-Simpsonian model to decompose λ'_P from Eqn 6 into contributions from the atmospheric window $\left(\lambda'^{w}_{P}\right)$ and water vapor absorption band $\left(\lambda'^{wv}_{P}\right)$:

$$\lambda'_P = \lambda'^{w}_{P} + \lambda'^{wv}_{P} \qquad (14)$$

where, following Jeevanjee (2018):

$$\lambda'^{w}_{P} = \int_{8\mu m}^{12\mu m} \pi \frac{dB\left(\lambda, T_s\right)}{dT_s} \mathrm{d}\lambda \approx 2Wm^{-2}K^{-1} \quad (15)$$

and

$$\lambda'^{wv}_P = \int_{\lambda \in (8\mu m, 12\mu m)} \pi \left(\frac{dB(\lambda, T_{em}(\lambda))}{dT_s} \right) d\lambda \approx 0 \quad (16)$$

assuming a partly-Simpsonian atmosphere such that $\frac{dT_{em}(\lambda)}{dT_s} \approx 0 \; for \; \lambda \notin (8\mu m, 12\mu m)$. This simplification provides an excellent approximation for the values of λ'_P simulated by GCMs (Ingram, 2013b; Zhang et al., 2020b). Errors arising from this approximation are largely the result of continuum absorption in the atmospheric window and pressure broadening of the water vapour absorption lines, both of which serve to make $\lambda'^{wv}_P < 0$ (Jeevanjee et al., 2021). This is largely offset by terrestrial emission within the water vapour bands from clouds and the surface in dry polar regions which causes $\lambda'^{wv}_P > 0$.

The second is that changes in terrestrial emission within the water vapour absorption bands are dominated by changes in relative humidity (*r*), not specific humidity or temperature (Möller, 1961). To a first approximation, $dT_{em}(\lambda) / d\ln(r) \approx -8K$ within the water vapour absorption bands (Soden and Bretherton, 1996), so every doubling of relative humidity results in roughly an 8K reduction in the water vapour emission temperature. This makes the climate system potentially quite sensitive to changes in relative humidity, particularly in the subtropical regions where *r* is small. However, theory, models and observations all support the relative invariance in *r* under climate change as discussed in the following section.

D. Changes in temperature and water vapour under global warming and their radiative impact.

1. The importance of different regions for water vapour and lapse rate feedback.

Key questions then, are what changes in humidity and temperature can we expect in a warming climate?

In the tropics, as the climate warms, increasing latent heat released within rising parcels, from their increased moisture content, leads to a steepening saturated adiabatic lapse rate (SALR)[9]. This means that temperature increases in the upper troposphere are greater than at the surface, increasing TOA OLR faster than implied by surface temperature change – a negative lapse rate feedback (Cubasch and Cess, 1990).

At the same time, warmer air can hold more moisture and the Clausius Clapeyron equation dictates that saturation specific humidity increases exponentially with warming. A key insight in the 1990s was that because of the opacity of the lower troposphere to LW radiation, and the increasing temperature contrast between the surface and atmosphere with height, *upper tropospheric humidity changes*, particularly in the tropics, have greatest radiative impact on TOA radiation change (Lindzen, 1990; Shine and Sinha, 1991; Rind and Lacis, 1993; Spencer and Braswell, 1997; Marsden and Valero, 2004; Inamdar et al., 2004). Furthermore, although for given specific humidity changes, the LW effects are greatest in the tropical lower troposphere (Colman, 2001), it is the *fractional* increase in specific humidity that determines the LW radiative impact. This is greatest in the upper troposphere if relative humidity does not change much with warming (Held and Soden, 2000).

Importantly, models suggest that relative humidity *does* remain close to unchanged, including in the upper troposphere, as the global temperature increases, thereby resulting in roughly exponential increases in saturation specific humidity with temperature under the Clausius-Clapeyron relationship. This implies a strong positive water

[9] The inability of the tropics to sustain strong tropospheric horizontal temperature gradients means that the tropical lapse rate, broadly, is set by the lapse rate in rising plumes within areas of active convection (Section IV-B). The expected slowing down of the tropical circulation under warming (Held and Soden, 2006) may reduce net convective mass flux, but not the temperature structure of the convection that does occur, and therefore would have little impact on broadscale lapse rate.

vapour feedback (Held and Soden, 2000). This means that the upper troposphere, particularly in the tropics and subtropics plays a disproportionate role in determining the global strength of LW water vapour feedback, as shown in Fig. 8.

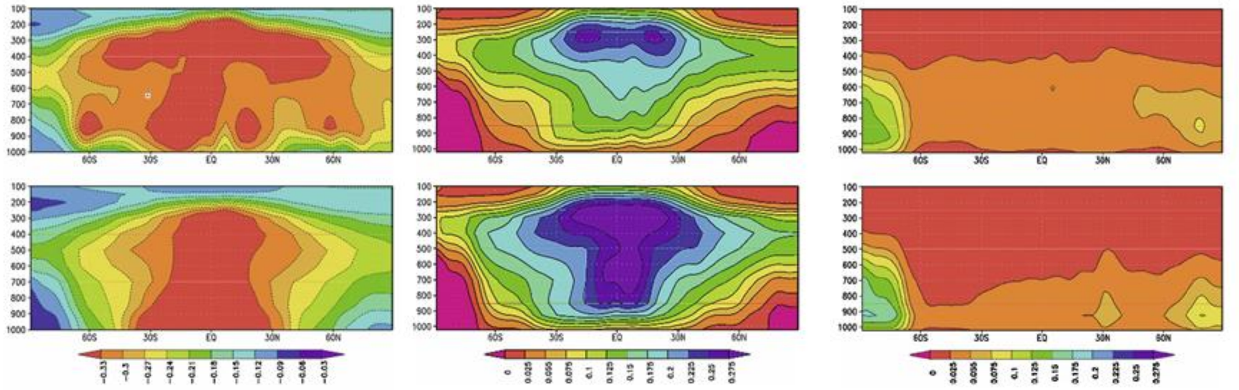


Figure 8: Zonal mean radiative "kernels", in height (hPa) and latitude (°), calculated using the Geophysical Fluids Dynamics Laboratory GCM. Shown is: (left column) LW TOA impact of 1K temperature increases at each point; (centre and right columns) LW and SW TOA impacts of moisture increases corresponding to a 1K temperature rise with fixed relative humidity. Top row shows "all sky" conditions, i.e. including the effect of clouds, bottom row "clear sky", i.e. with removal of clouds at all levels. Units are $Wm^{-2}K^{-1}100hPa^{-1}$. The importance of the tropical upper troposphere for LW water vapour is apparent, whereas low levels and high latitudes are most important for the SW. Source: Soden et al. (2008).

By contrast, in the SW, which contributes around 15% to global water vapour feedback (Colman and McAvaney, 1997) it is water vapour changes in the lower troposphere, and at high latitudes that are most important (Fig. 8). This is due to longer pathlengths from highly reflective surfaces/clouds and high summer insolation (Colman et al., 2001; Soden et al., 2008). Although the polar region SW feedback is consequently highly seasonal, compensation between hemispheres result in a fairly constant global SW feedback over the annual cycle (Colman, 2003b). In the LW the tropical dominance means the feedback also varies only weakly over the seasonal cycle (Colman, 2003b).

The geographical distribution of total *water vapour feedback* in a GCM is shown in Fig. 9b (Yoshimori et al., 2009). The dominance of the low latitude LW contributions can be seen, as can the presence of individual maxima off the equator, along with relatively low values from high latitudes, particularly in the southern hemisphere.

The distribution of *lapse rate feedback* for the same GCM is shown in Fig. 9a. Strong negative values are seen over the oceans throughout the tropics, consistent with convective changes. Positive areas occur over sea ice and high latitude land, corresponding to areas of low-level temperature inversions. The strength of lapse rate feedback in the tropics varies little with season, but large changes occur at high latitudes, again associated with the strength of the surface temperature inversion (Colman, 2003b). These issues will be discussed in more detail in Section IV-I.

In mid-latitudes, the presence of convective cores associated with baroclinic eddies also results in increased warming in the upper, compared to the lower, troposphere under global warming. This effect is stronger in the summer hemisphere than the winter hemisphere, and the Southern Hemisphere than the Northern (Frierson, 2006).

In summary, tropical and extra tropical regions contribute differently to both water vapour and lapse rate feedbacks. However, given the importance of the tropical upper troposphere for the globally dominant LW component of the water vapour feedback, many theoretical and observational studies over the past three decades have focused on understanding and evaluating humidity and temperature changes in this region.

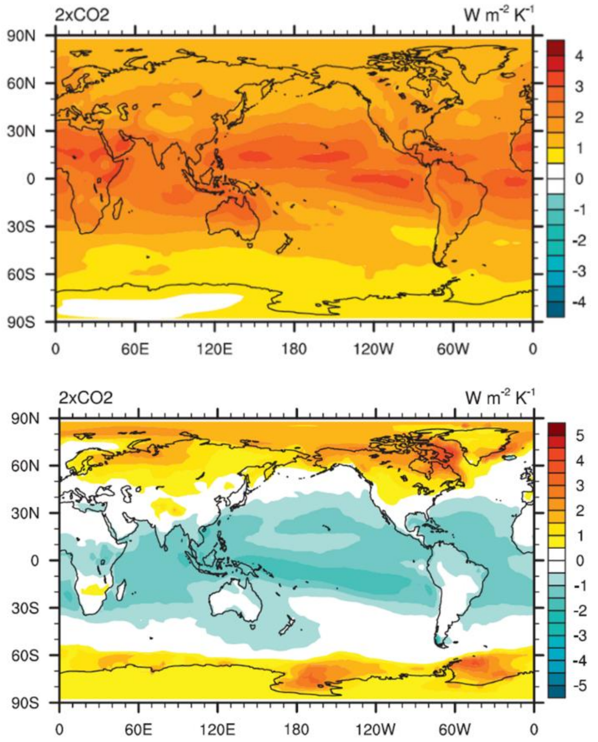


Figure 9. Geographical distributions of water vapour (top) and lapse rate (bottom) feedbacks under 2xCO_2 forcing, as calculated using the PRP methodology (see Appendix 1) applied to the CCSR/NIES/FRCGC/MIROC3.2(medres) Model for Interdisciplinary Research on Climate 3.2, medium-resolution version (Hasumi and Emori, 2004). Source: Yoshimori et al. (2009). © American Meteorological Society. Used with permission

2. Factors controlling relative humidity in a warming world.

Processes which may control humidity distribution in the atmosphere are potentially complex. These include detrainment from convective systems, cloud microphysical processes, including cloud droplet formation and re-evaporation, and turbulent mixing between clouds and ambient air, as well as large scale advective processes (Emanuel and Pierrehumbert, 1996; Emanuel and Zivkovic-Rothman, 1999), see Fig. 10. The large descending subtropical areas are particularly important for water vapour feedback because they are relatively cloud free and relative humidity is low, so they play a major role in radiation to space, and changes under warming therefore can have a large impact of global OLR (Pierrehumbert, 1995; Held and Soden 2000; Sherwood et al., 2010a,b). To establish the veracity of the feedbacks, particularly in the mid to upper troposphere, a combination of physical arguments and observational and modelling studies are needed.

3. Challenges to water vapour feedback "orthodoxy": convective drying and the role of microphysics.

Given the complexity of the tropics, and its importance for water vapour feedback, starting in the early 1990s some scientists raised challenges to the conventional role of water vapour feedback in climate change – that of it being a strong amplifying feedback. These challenges can be classed into four overall areas.

The first postulated that with the primary source of free tropospheric moisture being detrainment from deep convection in the tropics (see Fig. 10), deeper tropical convection in a warming climate could cause air to detrain from higher, colder regions, thereby resulting in strong decreases in relative humidity throughout regions of broadscale descent (Lindzen, 1990; Sun and Lindzen, 1993; Lindzen, 1994; Renno et al., 1994). This was postulated to lead to greatly weakened, or even negative water vapour feedback (Lindzen, 1990). A second challenge was that hypothesised large decreases in tropical high cloud fraction from convective outflows with warming could result in much drier air, primarily affecting high cloud cover but also reducing the strength of the water vapour feedback: the so-called "iris" effect (Lindzen et al., 2001). Related to this was a third conjecture that condensate outflows could decrease with warming, causing upper tropospheric drying. A fourth was that microphysical changes such as increased precipitation efficiency inside convective towers could reduce moisture supply to the upper troposphere. These "challenges" helped prompt intense research over the following two decades.

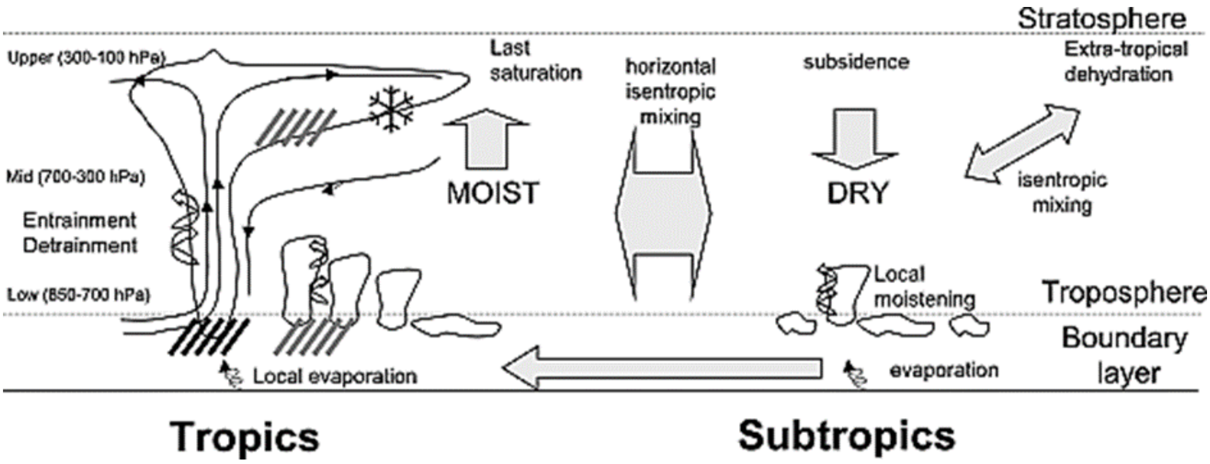


Figure 10. Schematic of key processes involved in moisture transport in the tropics and subtropics. Source: Sherwood et al. (2010b). Reproduced with permission from the American Geophysical Union.

Addressing the first point, the proposed simple model of detrainment from the tropical tropopause turns out to be too simplistic a view, with the atmosphere being moister than it would if all detrainment occurred at these temperatures (Held and Soden, 2000). In response to warming, observations show increased tropospheric temperature exceeding the cooling effect from higher detrainment, resulting in increased water vapour albeit with a small decrease in relative humidity (Minschwaner and Dessler, 2004; Minschwaner et al., 2006). A comparable, modest reduction in upper tropospheric relative humidity with warming has long been noted in climate models (Mitchell and Ingram, 1992; Held and Soden, 2000; Sherwood et al., 2010a; Fig. 11), consistent with a water vapour feedback around 5% weaker than implied by unchanged relative humidity (Soden and Held, 2006) and agreeing with observations within uncertainties (Minschwaner et al., 2006). This small reduction can be understood from temperature changes associated with convective outflow, in turn associated with the altitude of neutral buoyancy, and a consequence of the vertical gradient of longwave cooling associated with decreasing water vapour concentration with altitude driven by Clausius Clapeyron (Zelinka and Hartmann, 2012; Allan, 2012). Additionally, not all the air in the driest subtropical descent regions is sourced from deep tropical convection. These regions also contain air mixed in from mid-latitudes (Galewsky et al., 2005), indicating a role for dehydration in mid-latitude eddies as a source of subtropical dryness (Sherwood et al., 2010b).

In summary, the "convective drying" in a warmer climate proposed by Lindzen (1990) is now known to contribute only a very minor reduction below constant relative humidity.

On the question of an infrared "iris", there is no observational or modelling evidence of decreases in tropical high clouds of the magnitude proposed (22% per K of warming) (Chambers et al., 2002; Del Genio and Kovari, 2002; Hartmann and Michelsen, 2002; Lin et al., 2002, 2004; Rapp et al., 2005; Su et al., 2008). Nor is there evidence of significantly drier air resulting from decreased cloud cover (Fu et al., 2002). Just as with the first postulate ("convective drying"), the "iris" challenge was based in large part on simplified 2 box models of the tropics, and postulated relationships between moisture, detrainment and clouds not reproduced in GCMs or verified by observations.

The third and fourth challenges relate to the question of the role of cloud and convective "microphysics" in setting and changing upper tropospheric humidity under warming. The following section will address this important issue.

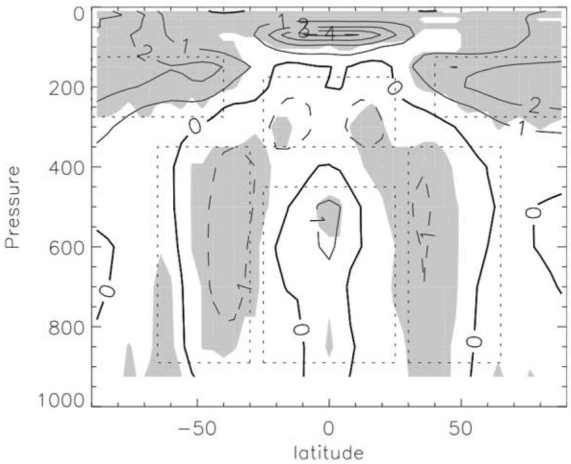


Figure 11: Ensemble mean change in relative humidity per K of surface warming under CO_2 forcing, calculated from 18 CMIP5 GCMs. Axes are height (hPa) versus latitude (°). Dashed contours indicate negative values, and shading represents areas of agreement on the sign of change by at least 90% of the models. Source: Sherwood et al. (2010a). Reproduced with permission from the American Geophysical Union.

4. The role of model parametrisations in humidity distributions

Although in the 1990s no GCM had shown weak or negative water vapour feedback (as indeed remains the case today), it remained possible that all GCMs were "wrong" in the same way, in other words all were missing or misrepresenting some key process. In particular, if details in so called "microphysics[10]", associated with convection and clouds are important for determining broadscale humidity distributions, then confidence in water vapour feedback would be substantially diminished considering the uncertainties in parameterisations of these processes in climate models (Randall et al., 2007; Boucher et al., 2013).

The reason why this question is so important is that confidence in models varies greatly for differing processes. It is high for the depiction of circulations of energy and moisture which are explicitly resolved by the model grid. GCMs used for climate modelling typically have grid sizes of around 50 to 100 km in the horizontal, and several hundred metres in the vertical[11]. However, many processes important for climate are not resolved on these scales, such as radiative interactions, convective plumes and entrainment, turbulent mixing and cloud droplet aggregation and precipitation processes. Such processes need to be "parameterised", that is represented in approximate form, based on relationships between the time and area averaged effects of the unresolved process and grid-resolved (i.e. "large scale") variables. Parameterisations may include processes that are difficult to measure or observe, or are based on empirical, theoretical, or statistically derived values (Mauritsen et al., 2012).

Uncertainties within parameterisations can have substantial impact on climate response. For example, experiments where a range of parameter settings are systematically changed within "realistic" (often empirical or expert assessed) ranges can sometimes produce very large differences in equilibrium climate sensitivity (Murphy et al., 2004; Stainforth et al., 2005; Collins et al., 2006; Klocke et al., 2011; Lambert et al., 2013; Tsushima et al., 2020). If water vapour feedback is sensitive to such parameter perturbations, in particular through control of the climatology or temperature dependency of upper tropospheric humidity, then confidence is reduced in the veracity of the feedback.

Evidence that parameterised microphysical processes are <u>not</u> critical for water vapour feedback come from several sources. The first is from "last saturation" simulations carried out using models without microphysics, showing that humidity distributions can be well represented using only evaporative and advective processes with 100% relative humidity limitation on parcels (Pierrehumbert and Rocca, 1998; Dessler and Sherwood, 2000; Gettelman et al. 2000; Sherwood

[10] "Microphysics" in this context refers to parametrised physical processes in cloud formation and convection, such as cloud droplet formation, coalescence and precipitation.

[11] Finer grid spacing of ~10 km or less in the horizontal may be achieved with embedded regional climate models, but computational limitations generally prevent such fine scale for long global experiments, and these models still require relevant physical parametrisations.

et al. 2006, 2010b). This is an important area of research in its own right and is discussed further in the next section.

The second is that strong positive water vapour feedback results from models with large numbers of different physical parameterisations and convection schemes (Colman and McAvaney, 1997; Ingram, 2002; Larson and Hartmann, 2003; Bony et al., 2006; Sanderson et al., 2010), as well as "cloud resolving" models[12] (CRMs, Thompkins and Craig, 1999) which have fewer unresolved convective processes than GCMs. For example, experiments using the Community Atmosphere Model (CAM4/5) suite of GCMs found that the magnitude of water vapour and lapse rate feedbacks were insensitive to a wide range of physical parameterisation changes beyond the representation of deep convection. These included changes to moist boundary layer and shallow convection schemes, stratiform cloud microphysics, aerosol impacts on cloud droplet formation and the model radiation code (Gettelman et al., 2012). Model climate sensitivity did change but was instead in response to changes in radiative forcing and tropical cloud feedbacks (Gettelman et al., 2012).

Very large (by a factor of 15) changes in detrainment related microphysics settings, in concert with other parameterization changes, *did* alter the magnitude of water vapour feedback in a large, perturbed parameter experiment by roughly ±12% (Sanderson et al., 2010). However, these setting changes were very large compared with the range of change commonly applied in GCMs for "tuning" purposes (Colman et al., 2018). Furthermore, there was strong offsetting from lapse rate feedback changes (Sanderson et al., 2010) and consistent with that another GCM showed little impact on climate sensitivity from convective detrainment changes (Mauritsen et al., 2012).

There have been suggestions that insufficient vertical resolution in GCMs means that sensitivity to microphysics may underrepresented (Emanuel and Zivkovic-Rothman, 1999; Tompkins and Emanuel, 2000). This has proven unfounded, however, as experiments show insensitivity of water vapour feedback to large changes in vertical resolution in GCMs (Ingram, 2002). Furthermore, water vapour feedback strength has not changed significantly over generations of models whereas vertical resolution has increased substantively – with models having up to ~100 layers in the vertical in the recent CMIP6 ensemble (Eyring et al., 2016; Voldoire et al., 2019).

Hints on the reasons for insensitivity of broadscale relative humidity distribution to microphysics come from studies in which atmospheric GCMs are forced by sea surface temperature (SST) and radiative perturbations to eliminate large-scale circulations such as the Hadley and Walker circulations and associated concentrated convective regions (Sherwood and Meyer, 2006). In this "boiling kettle" world, relative humidity in the upper troposphere was strongly affected by microphysical parameters determining precipitation efficiency. This was a factor which some had previously hypothesised might "rain out" extra moisture in the warmer world via convection, thereby decreasing relative humidity (Lindzen et al. 2001; Lau and Wu 2003). However, when convection was allowed to follow a more realistic, organised, structure, the sensitivity to precipitation efficiency microphysics in the model was strongly decreased in the same GCM (Sherwood and Meyer, 2006). The experiments found that upper tropospheric relative humidity sensitivity to doubling "convective precipitation efficiency" was only a few percent, implying that the presence of convective organisation makes the climate much less sensitive to the details of convective microphysics (Sherwood and Meyer, 2006).

The role of re-evaporation from cirrus clouds in determining upper tropospheric humidity has been examined using cloud observations from the International Satellite Cloud Climatology Project (ISCCP) and Television Infrared Observation Satellite (TIROS-N) Operational Vertical Sounder (TOVS) products and water vapour from combined Infrared/microwave retrievals (Luo and Rossow, 2004). Results show that, although cirrus can be an important sink of water vapour, its total water content is too small to have significant effect on

[12] Very high resolution models (down to around 10s of metres) capable of simulating individual convective clouds (Guichard and Couvreux, 2017).

upper tropospheric humidity through subsequent re-evaporation of condensate (Sherwood, 1999; Luo and Rossow, 2004; Soden, 2004; John and Soden, 2006). Instead, upper tropospheric moistening is associated with the same dynamical processes associated with the cirrus formation itself (Soden, 2004; Su et al., 2006).

Outside the tropics, parametrised microphysical processes are also not expected to be important for water vapour distribution. Vertical mixing in baroclinic eddies – which are explicitly resolved in GCMs – act to maintain relative humidity profiles at around 30-50% saturated throughout the year (Soden and Fu, 1995; Bates and Jackson, 1997; Stocker et al., 2001).

In summary, multiple lines of evidence show that neither water vapour distribution nor feedback are significantly sensitive to parameterisation choices in GCMs, including that of cloud or convection microphysics.

5. Simple models of water vapour distribution

Much understanding has been gained over the last two decades from application of the so-called "advection-condensation" (AC) approach. This idea poses perhaps the simplest possible explanation for humidity distribution within the atmosphere. It postulates that an air parcel's specific humidity is conserved, being set by its last saturation, then subject only to large-scale advection, with no moisture gains or losses through small-scale mixing, condensed water evaporation or further condensation (Pierrehumbert et al., 2007). Specific humidity is therefore set by last contact with the surface—the ultimate source of all atmospheric moisture—with subsequent losses due to vertical transport/convection resulting in cooling, condensation and precipitation from the parcel.

In the tropics the principal source of vertical advection is in convective regions associated with the upward branch of the Hadley circulation (Dessler and Minschwaner, 2007), with assumed saturation profiles—a feature supported by extensive observations (Bretherton et al., 2004; Holloway and Neelin, 2009). Outside the tropics final hydration is found to be sourced from air penetrating along isentropic surfaces to mid-latitudes (Dessler and Minschwaner, 2007). The AC approach explicitly excludes mixing on scales smaller than the resolved advective grid, and horizontal and vertical transport of condensed moisture (including processes such as re-evaporation of cloud droplets).

A large number of studies have followed, and conclude that large-scale moisture distributions are generally well represented by this simple framework (Sherwood, 1996a, b; Salathé and Hartmann, 1997; Pierrehumbert, 1998; Pierrehumbert and Roca, 1998; Dessler and Sherwood, 2000; Galewsky et al., 2005; Hurley and Galewsky, 2010), see Fig. 12. Both Eulerian and Lagrangian advective schemes have been used and results are not sensitive to this choice. Without diffusive processes such as turbulent mixing, the latter results in "filamentary structures" that increase in time, eventually necessitating some degree of spatial or temporal averaging (Sherwood et al., 2010b).

Theoretical studies have argued that the detrainment profile and subsequent humidity distribution of upper tropospheric moisture can be understood in terms of a straightforward balance between moistening from convective detrainment and large-scale clear-sky cooling and subsidence drying (Folkins et al., 2002; Folkins and Martin, 2005). Both the vertical structure of relative humidity and its tendency to be conserved as the surface warms can be reproduced by a simple analytical model using only the Clausius–Clapeyron relation, hydrostatic balance, and a bulk-plume water budget (Romps, 2014). This model provides an analytic expression for relative humidity, *r*,

$$r = \frac{\delta}{\delta + \gamma} \qquad (17)$$

where δ is the fractional detrainment, and $\gamma = -\,d\ln q^{*} / dz$ is the "water vapour lapse rate", with q^{*} the saturated specific humidity.

Relative humidity is large in the tropical upper troposphere where $\delta >> \gamma$ and convective moistening dominates over subsidence drying. Below this, r decreases due to increasing γ as one descends through the free troposphere. As the climate warms this simple analytical model predicts that δ and γ dependences on ambient temperature are roughly independent of surface temperature, implying a close to unchanged relative humidity profile (Jeevanjee, 2018).

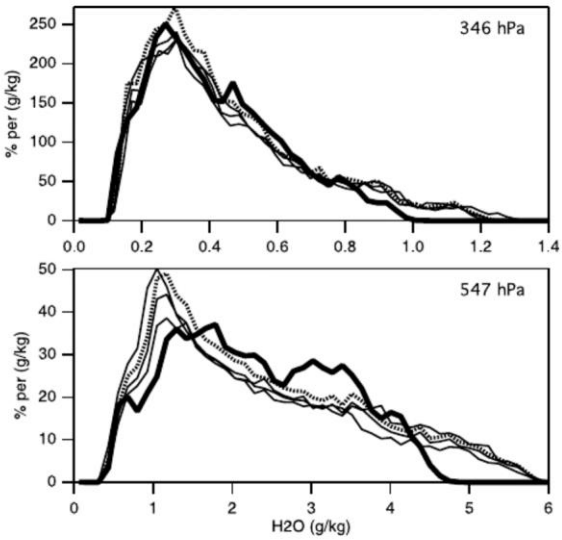


Figure 12: Results from an "advection-condensation" (AC) simulation of annual average water vapour mixing ratio (≈ specific humidity) at 346 hPa and 547hPa using 50 day moisture advection trajectories. Thick solid lines are AIRS satellite data. The three thin solid lines represent different configurations of the AC model, with different specified "microphysics", in this case being different "convective thresholds" whereby parcels mix with other sources of convection of different temperatures within rising plumes. Dashed lines show the AC model but with assumed relative humidity saturation limit of 90% instead of 100%.

The close overall agreement with moisture distribution between the AC model and observations is apparent, as well as the insensitivity of the AC model simulations to either "microphysics" specification, or even the precise definition of "saturation". Source: Dessler and Minschwaner (2007). Reproduced with permission from the American Geophysical Union.

Challenges in the AC approach include uncertainties in verifying observations both of moisture and winds in the mid to upper troposphere, leading to several different approaches. Validation of results has occurred against a number of satellite products sensitive to upper tropospheric humidity including 6.3μm HIRS brightness (e.g. Soden and Bretherton, 1996; Pierrehumbert and Roca, 1998), Microwave Limb Sounder (Dessler and Sherwood, 2000; Ryoo et al., 2009), AMSU-B (Brogniez and Pierrehumbert, 2006) and the Atmospheric Infrared Sounder (AIRS) (Dessler and Minschwaner, 2007).

AC studies have proved able to represent upper tropospheric relative humidity on a broad range of timescales including daily (Pierrehumbert and Roca, 1998), monthly (Dessler and Sherwood, 2000) and annual. The approach has also been found to skilfully represent GCM moisture fields using model winds (Salathé and Hartmann, 2000; Galewski et al., 2005 [check this 1]). Nor is the skill adversely affected by the imposition of convective microphysical assumptions, or even to the level of relative humidity designated as "saturated" – e.g. reducing from 100% to 90% to account for mixed detrainment below relative humidities of 100% (Dessler and Minschwaner, 2007; Fig. 12). An extensive review of the AC approach is provided by Sherwood et al. (2010b).

In summary, then, AC simulations can skilfully reproduce large-scale tropical humidity (particularly in the upper troposphere), typically to within 10% accuracy over a very wide range of humidity levels and regimes (Sherwood et al, 2010b; Fig. 12). The reason they can do this appears to be because (i) at the large-scale, free atmospheric parcel source regions of moisture can be effectively described as at or near saturation and (ii) additional sources of moisture such as condensed cloud water or ice are modest over parcel lifetime to next saturation (Sherwood et al., 2010b).

The key importance of theoretical and AC studies is they provide convincing evidence that convective, or cloud microphysical settings have little impact on humidity distribution, meaning that it is unlikely these uncertainties substantially affect projected humidity changes under a warming climate. The only remaining possible sensitivity would be on the advecting *winds* themselves (Dessler and Minschwaner, 2007). As succinctly stated by Dessler and Sherwood (2000): "We see no evidence to suggest that accurate predictions of the humidity in this (upper tropospheric) region are dependent on accurate simulations of microphysical processes or on transport of ice or liquid water. Our results instead suggest that accurate predictions of the humidity primarily require realistic three-dimensional large- scale (greater than a few hundred kilometers) wind fields."

6. Deviations from unchanged relative humidity.

Climate models project that on broad scales relative humidity is close to unchanged throughout much of the troposphere under global warming. This is not exact however, and some systematic large-scale deviations from uniformity are apparent, as shown in Fig. 11 for the multi-model mean for 18 CMIP5 models (note that this is for equilibrium change). These deviations are important to understand in that they clarify model processes controlling moisture distribution, have a modest effect on the strength of the global (LW) feedback, and have a critical role in contributing to cloud feedback (Ceppi et al., 2017; Sherwood et al., 2020).

As the climate warms, models predict (Gettelman et al., 2010; O'Gorman and Singh, 2013), and observations confirm (Santer et al., 2003a,b) that there is an increase in the height of the tropopause. This is a consequence of the decreased effectiveness of thermal emission from water vapour below around 200 K (Hartmann and Larson, 2002). This increases relative humidity in the region of both the tropical and extratropical tropopause – i.e. at heights where formerly dry stratospheric air is replaced by moister tropospheric conditions (Boucher et al., 2013). Both clouds and relative humidity shift upward following a fixed temperature coordinate (Po-Chedley et al., 2019), consistent with that expected from a fixed temperature for convectively detrained clouds and moisture (Hartmann and Larson, 2002; Romps, 2014). Maximum zonal mean values of these changes are around 2-4% of relative humidity per K of warming (Fig. 11). This has important consequences for the increasing strength of water vapour feedback in warmer base climates (Section IV-G).

Other regions of increasing relative humidity are in the equatorial tropics (below about 400hPa) and at high latitudes throughout the depth of the atmosphere. Decreases occur in the upper troposphere in the tropics, and through a broad depth in the sub-tropics to mid-latitudes. Notably, there is a marked symmetry between both hemispheres, indicating that the circulation changes resulting in relative humidity perturbations are not sensitive to continental distribution (Sherwood et al., 2010a). Furthermore, the presence of modest increases and decreases results in close-to-unchanged relative humidity globally as temperature rise.

What causes these large-scale humidity changes? The simplest possible explanation – the so-called "shift" hypothesis, postulates that they result from upward and poleward expansion of tropical circulations – i.e., decreases are located where relative humidity in the current climate increases either with altitude, as it does in the tropical upper troposphere, or with latitude, as it does in mid-latitudes (Sherwood et al., 2010a). Sherwood et al., 2010a however showed that mid-latitude humidity changes are 2 to 3 times too large for this simple explanation. Instead, they postulate that air parcels in drying regions last experience saturation in regions of the atmosphere warming at a relatively slower rate – i.e. resulting from nonuniform rates of warming or wind change (Hurley and Galewski, 2010).

These relative humidity differences are modest on global scales. Therefore, although they are very important for cloud feedbacks, they have only a small impact on global water vapour feedback (Boucher et al., 2013). Models simulate water vapour feedback of around 5% weaker than that predicted by fixed relative humidity (Soden and Held, 2006; Soden et al., 2008; Vial et al., 2013; Colman and Hanson, 2016) principally as a result of the reduction in upper tropospheric relative humidity (Vial et al., 2013). High-resolution CRMs also find upward shifts in relative humidity with increasing temperature (Kuang and Hartmann, 2007) adding confidence to GCM processes.

In summary then, small changes in relative humidity are found in models under global warming, partly caused by processes such as a rising tropopause and non-uniform rates of warming. Their net effect on water vapour feedback is to reduce it in strength by around 5%.

7. Water vapour feedback and convective aggregation

As the world warms, changes in "aggregation" (clustering) of tropical convection (e.g., Muller and Held, 2012) could potentially affect water vapour feedback strength. This is because changes in the way convection is organised may affect broadscale humidity, and in particular the area or dryness of large-scale descending regions. These large, relatively cloud free areas play a major role in global radiation balance (Pierrhumbert, 1999; Peters and Bretherton, 2005). Observations of changes in convective self-aggregation do suggest an anti-correlation between aggregation and tropospheric humidity outside the boundary layer (Tobin et al., 2012; 2013). Furthermore, drying of the free troposphere during periods of greater aggregation has been found to increase the clear-sky OLR over the tropics and constitute the dominant factor controlling interannual variability of the tropical-mean radiation budget (Bony et al, 2020).

A recent, large, multi-model ensemble using models ranging from GCMs to CRMs, the Radiative- Convective Equilibrium Model Intercomparison Project (Wing et al., 2018), found models widely exhibited self-aggregation, which acted to warm and dry the troposphere in the current climate (Wing et al., 2020). Under *global warming* however there was no clear tendency in the degree of self-aggregation, although there was some sensitivity in this result to the use of parameterised (GCM) convection versus CRMs (Wing et al., 2020).

There remains some uncertainty, then, on how convective aggregation may change in response to a warming climate and therefore what the net effect would be on water vapour feedback.

Recently a convection related negative "water vapour buoyancy" feedback has been proposed, whereby the relative lightness of water vapour compared with other atmospheric constituents induces increased buoyancy in moist regions, compensated by increased temperatures in dry regions (Seidel and Yang, 2020). This is hypothesised to lead to increased OLR, and to strengthen with warming, producing a negative feedback (Seidel and Yang, 2020). However, climate models explicitly represent the density impact of moisture and its effect on circulations, so this represents merely a different way of subdividing known feedback processes, rather than a new negative feedback.

8. Impact of water vapour or lapse rate changes on radiative forcing.

Apart from temperature related changes (Eqn 1) the question arises whether rapid water vapour or lapse rate adjustments in response to *the forcing itself* (such as from a sudden increase in CO_2) induce an additional TOA impact. Such "rapid responses" (i.e. fast, non-surface temperature related atmospheric adjustments) are very important for clouds (Gregory and Webb, 2008; Sherwood et al., 2015), and are regarded as contributing to the initial forcing.

However, there is little evidence on a *global* scale of rapid, radiatively-important, water vapour adjustments following CO_2 forcing (Colman and McAvaney, 2011; Block and Mauritsen, 2013; Vial et al. 2013; Po-Chedley et al., 2018). On a *regional* scale, positive and negative radiative responses can occur (Block and Mauritsen, 2013), although evidence of rapid responses over land has been linked to fast land warming, rather than atmospheric adjustments (Vial et al., 2013). Together these studies mean that to a very good approximation water vapour acts as a true "feedback process" coupled with global surface temperature change following external radiative forcing, and contributes little to the original forcing.

Related to this issue, questions have also been raised as to whether human activities such as irrigation directly affect water vapour concentrations in the atmosphere, with consequent radiative impact. Idealised simulations by Boucher et al. (2004) found indeed water vapour concentrations are increased locally in irrigated regions, resulting in global mean OLR impact of between 0.03 and 0.1 Wm^{-2}. This of course is a forcing process, not a feedback. However, it is unclear if this is even truly a radiative forcing process, as local surface temperatures were decreased, rather than increased, in their experiments due to surface evaporative and other changes (Boucher et al., 2004). Furthermore, vapour lifetime in the atmosphere is short, around two weeks (Zhang et al., 2003; Lacis et al., 2013; Stevens and Bony, 2013), meaning that without increased temperatures injected water vapour will rapidly precipitate out.

Another source of direct water vapour injection, that from aviation, is estimated around an order of magnitude less impact again than that from irrigation (IPCC, 1999), and so makes negligible contribution to global radiative forcing.

Lapse rate response can be sensitive to the forcing agent due to differences in vertical forcing, and resultant rapid response differences in the large-scale tropospheric stability (Ceppi and Gregory, 2019). This includes changes in upper tropospheric temperature (Andrews and Forster, 2008). The global radiative impact on forcing for CO_2 increases, however, is small (Colman and McAvaney, 2011; Vial et al., 2013).

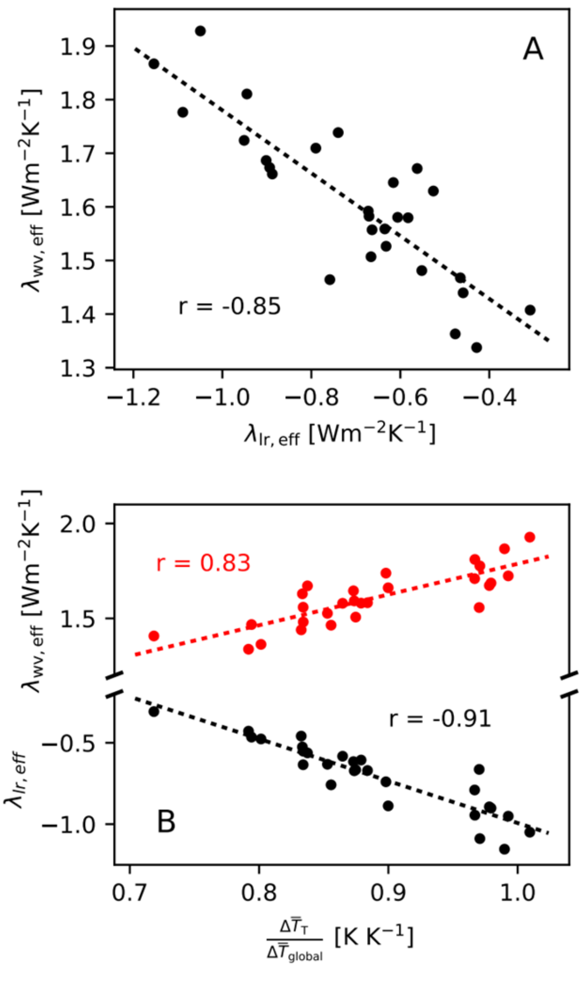


Figure 13. (Colour online) (a) The relationship between global "effective" (viz standardly defined) water vapour and lapse rate feedbacks for CMIP5 models. Each data point represents a GCM, and the correlation coefficient is shown. (b) Water vapour and lapse rate feedbacks plotted against the ratio of tropical to global surface temperature change. Note the discontinuity on the y-axis. Source: Po-Chedley et al. (2018).

This shows the strong anticorrelation between water vapour and lapse rate feedbacks golbally, and the close dependence of lapse rate feedback on the relative warming between the tropics and extratropics.

E. Relationship between water vapour and lapse rate feedbacks

1. What causes the (anti-) correlation between water vapour and lapse rate feedbacks?

It has long been noted that there is a strong offsetting relationship between the traditionally defined water vapour and lapse rate feedbacks leading to a reduced overall range of feedbacks across multi-model ensembles (Cess, 1975; Colman, 2003a; Soden and Held, 2006; Held and Shell, 2012; Koll and Cronin, 2018). This occurs not just across different models, but also within individual models when modifications such as parameterisation changes are made (Zhang et al., 1994). For example, a weakened lapse rate feedback was largely offset by a corresponding weakened water vapour feedback when moving from one version of the model to another of the Community Climate System Model (CCSM) (Bitz et al., 2012), or when adjusting parameters as part of a large "perturbed parameter" ensemble within a single model (Sanderson et al., 2010). Figure 13a (Po-Chedley et al., 2018) illustrates offsetting across Coupled Model Intercomparison Project, phase 5 (CMIP5, Taylor et al., 2012) GCMs.

Understanding of this anti-correlation has evolved significantly over time. Early focus emphasised the tropical upper troposphere, given its critical role in determining water vapour feedback strength. Enhanced warming in this region it was argued, produces a stronger negative global lapse rate feedback, but also strengthens water vapour feedback, due to consequent increased specific humidity under the additional warming (Cess, 1975; Randall et al., 2007; Huybers, 2010). However, realisation that both global water vapour (Soden and Held, 2006) and lapse rate feedback strength (Shell, 2013) are correlated with equator to pole temperature gradients hinted at another mechanism. Consideration of regional feedbacks (Armour et al, 2013), found that correlation between water vapour and lapse rate feedback in the tropics was in fact weak (Fig. 14a). Furthermore, inter-model variation in the tropical water vapour feedback strength could be largely explained by changes in relative humidity, rather than resulting from tropospheric temperature changes under a fixed relative humidity assumption (Vial et al., 2013; Po-Chedley et al., 2018). This is illustrated in Fig. 14b, which shows strong correlation between the classically defined tropical water vapour feedback λ_q (Eqn 4) and the Held and Shell (2012) relative humidity term λ'_H (Eqn 8). On the other hand, no link is found between changes in relative humidity and changes in tropical lapse rate feedback (Po-Chedley et al., 2018). Hence no common physical mechanism links variations in tropical (mean) water vapour and lapse rate feedback strengths, so little correlation would be expected (or is found – Fig 14a).

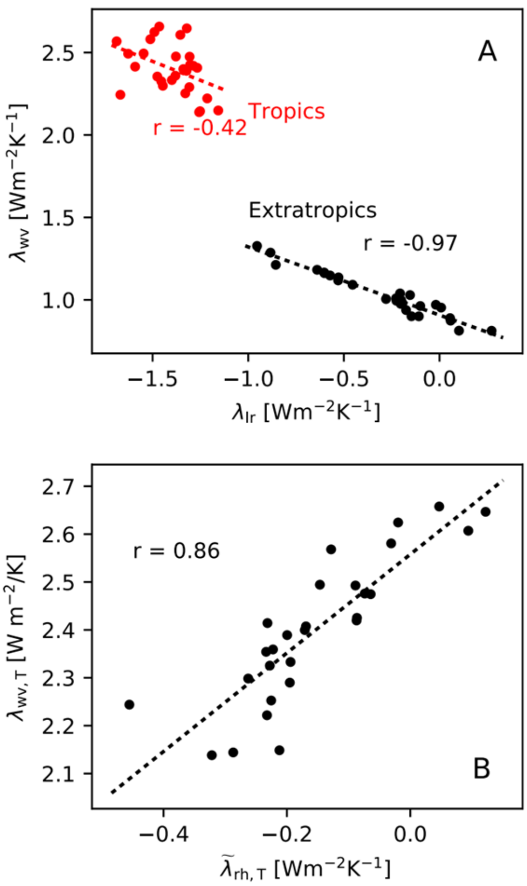


Figure 14. (Colour online) (a) The relationship between global "effective" (viz standardly defined) water vapour and lapse rate feedbacks for the CMIP5 models in the tropics (latitude < 30°) and extra tropics. (b) Tropical water vapour feedback shown against tropical relative humidity feedback (λ'_H in Eqn 8), here denoted as $\tilde{\lambda}_{rh}$. Source: Po-Chedley et al. (2018). © American Meteorological Society. Used with permission

By contrast, in the extra tropics very strong anti-correlation is found between classically defined water vapour and lapse rate feedback, as shown in Fig. 14a. Importantly, the *range* in lapse rate feedback is greater in the Southern Hemisphere than the northern by a factor of three. Po-Chedley et al. (2018) showed that the asymmetry was due to contrasts in the spread in temperature pattern differences across models. Throughout the northern hemisphere, surface and atmospheric temperature changes are strongly coupled with the tropics. In the Southern extra-tropics, by contrast, temperature correlation with tropical changes is relatively weak, and instead dominated by patterns of surface temperature change and consequent local feedbacks (Po-Chedley et al., 2018). Moreover, delayed warming in the Southern Hemisphere extratropical latitudes results in regional feedbacks that are sensitive to poleward mixed warming and moistening from the tropics (Butler et al., 2010; Rose and Rencurrel, 2016). It is differences, then, in both the magnitude and pattern of Southern Hemisphere extratropical warming, in turn related to differences such as Antarctic sea ice climatology (Feldl et al., 2017b) that drive the inter-model range of feedback strengths, and the global anti-correlation seen in Fig. 13a.

Consistent with this, for the combined water vapour plus lapse rate feedback, a recent study of 31 CMIP5 models showed that relative humidity differences at the regional level contribute around 40% of the inter-model variance, in turn coupled to differences in patterns of tropical SST changes, with the remainder scaling closely with the difference between tropical/subtropical mean temperature change and the extra tropics (Zhang et al., 2020a). This highlights the utility of the alternative RH-based feedback framework (Section III-B) for understanding the cause of intermodel spread in climate sensitivity.

Paleo studies reinforce the extratropical importance for the offsetting nature of water vapour and lapse rate feedbacks. In Last Glacial Maximum (LGM, 19-27kyr BP) experiments, northern hemisphere continental ice sheets push cold surfaces to much lower latitudes, weakening water vapour feedback, but also rendering lapse rate feedback weakly positive globally, resulting in virtually unchanged global water vapour plus lapse rate feedback (Yoshimori et al., 2009).

The above discussion relates to climate change timescale feedbacks. On shorter timescales, too, anticorrelations can be found in the feedbacks. Evaluation of global lapse rate and water vapour feedbacks using accurate so-called "partial radiation perturbation" (PRP) methods (see Appendix 1) every six hours for six consecutive years of model simulations show only weak correlation on six hourly timescales, but strong anti-correlation between the feedbacks on three monthly (i.e. seasonal) timescales (r=0.71) (Klocke et al., 2013). Different models have been found to produce similar *interannual* fluctuations in clear sky (i.e. cloud free) OLR with surface temperature fluctuation when forced with observed SST changes, showing that their combined water vapour plus lapse rate feedbacks were similar, although both lapse rate changes and moisture distributions differed strongly, with very different upper tropospheric warming, suggesting contrasting mechanisms (Allan et al., 2002).

Regionally however, there is little *geographical* (anti-)correlation between the strength of water vapour and lapse rate feedbacks in either models (Taylor et al., 2011) or observations (Ferraro et al., 2015). This can be understood from the differing processes driving these feedbacks at regional or local scales. The close to ubiquitous saturated adiabatic lapse rate in the tropics, and weak horizontal temperatures gradients above about 700hPa, means that local *lapse rate* feedback variations are primarily driven by surface temperature change patterns – stronger reduced OLR corresponds to greatest temperature increases (Lambert and Taylor, 2014). Local *water vapour* feedback on the other hand, is less tied to surface temperature increases. Over land, where surface temperature increases are greatest, relative humidity in the lower part of the atmosphere decreases because of reduced moisture availability (Fasullo, 2010). This modestly weakens local water vapour feedback (Lambert and Taylor, 2014). Over oceans, local maxima in water vapour feedback are related only weakly to surface temperature changes and are instead strongly associated with areas of increased heavy precipitation (Lambert and Taylor, 2014). In summary then, because of the differing processes driving *regional* changes in water vapour and lapse rate feedbacks, little (anti-) correlation is found at these scales.

2. What causes the spread in combined water vapour + lapse rate feedbacks in models?

The previous sub-section discussed the reasons for the anti-correlation between separate water vapour and lapse rate feedbacks in GCMs. There is now much better understanding of the source of spread in the *combined* feedback. This is an important issue, as it is the combined feedback that underpins overall climate sensitivity and the uncertainty of the combined feedbacks that contributes to sensitivity spread (Fig. 3). Furthermore, the cause of the spread casts light on the spread that remains under the alternate RH based formulation of feedbacks described in Section III-B. The spread results from different processes at different latitudes.

In the tropics, the offsetting upper tropospheric contributions from water vapour and lapse rate changes ensure that range in the combined feedback relates primarily to relative humidity changes, rather than changes to the vertical structure of temperature or moisture (Held and Shell, 2012; Vial et al., 2013; Po-Chedley et al. 2018). Theoretical support also comes from the realisation that relative humidity changes alone are important for determining the combined feedback strength, provided that infrared absorption bands are close to saturated (Ingram, 2010, 2013a, b), as they are throughout much of the atmosphere (Section IV-A).

The reasons for *tropics-wide* differences in relative humidity changes in models are not fully understood, but likely relate to differences in SST warming patterns (Andrews and Webb, 2018; Armour et al., 2013) or parametrisation differences to processes such as deep convection (Po-Chedley et al. 2018). Consistent with this, a recent perturbed physics ensemble found that convective parameters determining the entrainment of environmental air into convective plumes controlled present-day climate clear sky TOA LW fluxes (representing the clear sky combination of Planck, water vapour and lapse rate feedbacks), as well as their response to global warming (Tsushima et al., 2020). Furthermore, there was a strong relationship between current climate tropical mean clear sky OLR and flux change (Tsushima et al., 2020).

In mid latitudes, relative humidity changes are not related to changes in the net feedback, and models show little range in relative humidity changes over the poles (Vial et al., 2013). Instead, the spread in the combined water vapour+lapse rate feedback in the mid-latitude and polar regions depends on spread in the lapse rate feedback (Vial et al., 2013). As discussed in the previous section, then, the spread in absolute feedback strength in the *extra tropics* depends upon local feedbacks in those regions (Po-Chedley et al. 2018).

Clouds differences may also have impact on combined feedback strength. The presence of clouds compared to "clear sky" conditions has different impacts on LW and SW components of the water vapour feedback. In the smaller SW component, feedback is modestly strengthened (Zhang et al., 1994; Fig. 8), largely due to increased path length from multiple reflections (Colman et al., 2001; Soden et al., 2008). For the dominant LW component, feedback is weakened (Fig. 8), due to mid and upper-level clouds partially obscuring the TOA radiative impacts of underlying water vapour changes (Soden et al., 2004). On the other hand, for lapse rate feedback, the presence of upper tropospheric clouds strengthens the impact of upper tropospheric temperature changes, because clouds are stronger infrared emitters and absorbers than clear sky (Zhang et al. 1994).

Intriguingly, a covariance has been noted between CMIP3 model combined water vapour plus lapse rate and cloud feedbacks (Huybers, 2010) with stronger feedback in one implying weaker in the other. Although it remains possible that this is an artefact of feedback evaluation methodology/statistics, it may be related to physical processes, such as convective differences between models resulting in different changes in tropical upper tropospheric relative humidity, with compensations between water vapour feedback and changes in anvil cloud cover (Huybers, 2010). "Suppressed feedback" experiments in one GCM have also noted interactions between water vapour, lapse rate and cloud feedbacks (Mauritsen et al., 2013). Extra upper tropospheric warming due to increased cloud, strengthened water vapour feedback, but at the same time rising convective cloud strengthened negative tropical lapse rate feedback, damping the warming from cloud changes (Mauritsen et al., 2013). These processes and correlations have only had limited attention however, and further research would be needed to understand their significance for net feedback strength and climate change.

In summary, the source of inter model spread in combined (traditionally defined) water vapour-lapse rate feedback is from tropics-wide relative humidity differences in low latitudes, and from lapse rate feedback spread in the extra-tropics. Under the RH based formulation (Section III-B), the tropical effect in included in the separate "relative humidity feedback" term showing there are benefits of this approach in identifying sources of inter-model feedback spread. Differences in cloud cover between models may also play a role in combined feedback spread, and there remain hints of correlations with cloud feedbacks that are not confirmed or fully understood.

F. Stratospheric water vapour feedback

Traditionally, water vapour feedback was perceived as confined to the troposphere, albeit with increasing tropopause height implying higher level contributions in a warmer climate (Santer et al., 2003a, b; Meraner et al., 2013).

Enhanced climate *forcing* (a non-feedback process) can occur from stratospheric methane oxidation (Forster and Shine, 1999; Forster et al. 2007). Furthermore, model results suggest stratospheric water vapour changes can amplify forcing from increases in lower stratospheric ozone. This occurs through increased stratospheric water vapour inducing a "secondary forcing" (Stuber et al., 2001). Stratospheric water vapour adjustments however, have negligible impact on CO_2 forcing, or ozone forcing in the troposphere (Stuber et al., 2001).

At first glance, however, we might expect relatively little role for the stratosphere in water vapour *feedback*. In contrast to the troposphere, there are no reasons, *a priori*, to expect, say, unchanged relative humidity in the stratosphere (Stuber et al., 2001), and fractional changes in stratospheric water vapour have less impact radiatively than do those of the upper troposphere (Allan et al., 1999).

Observations suggest that stratospheric increases in water vapour have affected TOA radiation over recent decades (Solomon et al., 2010) and examination of satellite and reanalyses data link lower stratospheric water vapour changes with surface temperature changes suggesting it is operating as a true feedback (Dessler et al., 2013). It has long been thought that lower stratospheric water vapour enters through the tropical tropopause, with amounts controlled by minimum tropopause temperatures (Brewer, 1949; Rosenlof et al., 1997; Joshi and Shine, 2003). Processes involved are a combination of convective and broad-scale ascent (Keith, 2000; Sherwood and Dessler, 2000; Rosenlof, 2003). However, recent evidence reveals a broader picture, with water vapour entering the stratosphere both through the tropical and extratropical tropopause (Dessler et al., 1995), due to tropopause warming offsetting the "freeze-drying" process (Gettelman et al., 2009; Smalley et al, 2017). Models also suggest that for very strong warming (e.g. under $8xCO_2$), large upper tropospheric warming greatly reduces the tropopause "cold trap", leading to enhanced penetration of water into the lower stratosphere (Russell et al., 2013; Lacis et al., 2013).

CMIP5 models robustly show stratospheric moistening with global warming (Gettelman et al., 2010; Smalley et al., 2017). This produces significant additional radiative perturbations, peaking in mid-latitudes, with most of the contribution – over three quarters – resulting from extratropical lower stratospheric processes (Banerjee et al., 2019). Estimate of the strength of the associated feedback is 0.15±0.04 $Wm^{-2}K^{-1}$, with a range of 0.10–0.26 $Wm^{-2}K^{-1}$ (Banerjee et al., 2019). The upper end of this multi-model estimate is comparable to the previous single model estimates of Dessler et al. (2013) and Stuber et al. (2001). A much lower CMIP5 multi-model estimate of 0.02 ± 0.01 $Wm^{-2}K^{-1}$ using a similar kernels-based methodology[13] (Huang et al., 2016) appears unreliable because of the use of "instantaneous" radiative kernels, that is ones which do not consider stratospheric temperature adjustments (Solomon et al., 2010; Maycock and Shine, 2012; Banerjee et al., 2019).

Recent work has aimed to understand processes better, and limitations or biases in models. It is important to note that some potentially important troposphere/stratosphere exchange processes appear to be underdone by models, such as the effect of the quasi-biennial oscillation (QBO) on humidity in the lower stratosphere (Smalley et al., 2017). This may be significant for the strength of the feedback, as the QBO plays an important role in modulating stratospheric water vapour through its effect on tropical tropopause temperature (Tian et al., 2019). Ozone effects on stratospheric water vapour feedback are also not well understood; experiments with partially prescribed rather than fully interactive ozone had the effect of increasing climate sensitivity through modification of the

[13] See Appendix 1 for description

water vapour feedback strength (Nowack et al., 2018).

Nor is the importance of the additional radiative effects of stratospheric water vapour feedback fully established. For example, it is postulated that there are several radiation relevant processes from the increased stratospheric water vapour in a warmer climate. There is direct suppression of OLR from the additional atmospheric radiative opacity. Additionally, the stratosphere cools from the combined radiative effect of additional water vapour in both the troposphere and the stratosphere, which further reduces OLR (Wang and Huang, 2020). Offsetting this, ongoing tropospheric warming provides additional upwelling LW radiation, inducing stratospheric warming and increased OLR. The combined result may then result in negligible net TOA radiative flux changes (Wang and Huang, 2020). A recent stratospheric water vapour "locking" experiment indeed found only a 2% increase in surface warming under $4xCO_2$ forcing (Huang et al., 2020) due to warming from increased stratospheric water vapour being compensated by cooling from upper troposphere moisture and cloud responses. However, a similar experiment with a chemistry-climate model found a stratospheric water vapour feedback of 0.11 $Wm^{-2}K^{-1}$, contributing around 10% to global warming under CO_2 quadrupling (Li and Newman, 2020).

The science remains unsettled in this area. It is clear from examination of the CMIP5 ensemble that there is significant LW impact of increased lower stratospheric water vapour with warming, and observational evidence indicates recent impacts on TOA radiation. However, evidence from single model experiments also suggests that compensating negative feedbacks in the troposphere from clouds or temperature changes may result in negligible *net* enhancement of global warming, although the results are somewhat inconsistent. Further research is needed to untangle and quantify these effects, particularly sampling the results from multi-model ensembles.

G. State and forcing dependence of water vapour and lapse rate feedbacks.

Much of the research to date has focused on vapour and lapse rate feedbacks in the "current" climate. However, the climate is continually changing, and in recent years it has been demonstrated that there is a marked state dependency of climate sensitivity, and consequently the concept of fixed strength feedbacks needs to be revisited (Knutti and Rugenstein, 2015; Rugenstein et al., 2020). Indeed, the general paradigm of considering changing "equilibrium" feedbacks can overlook important dynamic and timescale components of their response (Hallegate et al., 2006).

Water vapour and lapse rate feedbacks have been found to evolve on timescales from decades to centuries, as GCMs slowly equilibrate in response to an impulsive doubling or quadrupling of CO_2 (Armour et al., 2013). This evolution in turn is the result not just of global mean temperature change, but also of changing meridional surface temperature patterns, such as delayed warming in the Southern Ocean and southern high latitudes, which affect the balance between low and high latitude feedback contributions (Armour et al., 2013; Shell, 2013; Andrews et al., 2015; Dessler, 2020). On extremely long (millennial) time scales, GCM clear sky LW feedback (a combination of Planck, lapse rate and water vapour) becomes steadily less stabilising. This is sourced mainly in the tropics and northern hemisphere mid-latitudes, consistent with strengthening water vapour feedback and an increasing tropopause height (Rugenstein et al., 2019).

Paleo climates provide the opportunity to test water vapour and lapse rate feedback under different base conditions, and under different forcing. For example, during the LGM CO_2 concentrations were around 2/3 preindustrial levels (along with reduced amounts of other GHGs), and there was additional forcing from vegetation changes and extensive ice sheet coverage of northern continents (Masson-Delmotte et al., 2013).

State dependency is apparent across different paleo regimes (Berger et al., 1993; Crucifix, 2006; Lariviere et al., 2012), with positive feedbacks overall becoming stronger as the climate warms. In a modelling study comparing modern day and early Paleogene (~65-35 million years ago) in which global mean temperature changes by 12° C, this strengthening was attributed largely to cloud feedback, as increasingly strong water vapour feedback was close to offset by increasingly negative lapse rate feedback (Caballero and Huber, 2013).

A GCM experiment with current day surface properties but LGM-level CO_2, CH_4 and N_2O, reveals little difference in the strength of water vapour and lapse rate feedbacks (Yoshimori et al., 2009; Fig. 15). Adding LGM ice sheets does not change SW water vapour feedback, but has a profound effect in the LW, reducing it by around 25%, and overall feedback by roughly 22% (see Fig. 15). This is largely because resultant extratropical temperature changes are much greater than tropical, and these are regions of relatively weak water vapour feedback (Yoshimori et al., 2009). Lapse rate feedback in the LGM experiment is globally *positive* (see Fig. 15), because cold, ice covered surfaces extend to relatively low latitudes, causing positive lapse rate feedback in these regions to outweigh negative tropical contributions (Yoshimori et al., 2009). As found in many other scenarios, offsetting changes in water vapour and lapse rate feedback result in a close to unchanged combined feedback (Fig. 15). This offsetting means that a RH based formulation of the feedbacks would suggest little change to the Planck or "lapse rate" terms across all experiments.

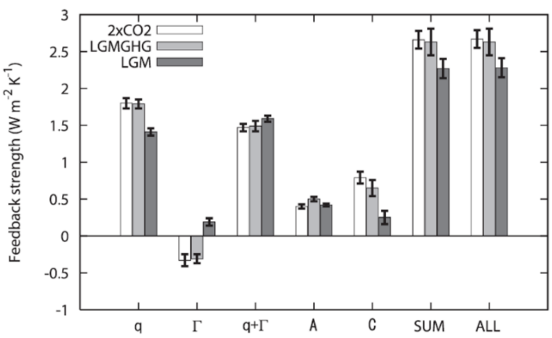


Figure 15. Feedbacks derived from the Model for Interdisciplinary Research on Climate 3.2, medium-resolution version [MIROC3.2(medres)] GCM under $2xCO_2$ forcing above current climate, as well from Last Glacial Maximum (LGM) changes compared with the current climate. LGM forcing includes ice sheets over northern continents and CO_2 levels of 185 ppm (~65% of current value) along with reduced values of CH_4 and N_2O. LGMGHG denotes an experiment with current climate ice sheets, but with LGM level greenhouse gases. Feedback notation is as in Eqn 4, except for "A" as surface albedo. Error bars represent one standard deviation of different 10-year samples, calculated using PRP (see Appendix 1).

This shows the weakened water vapour feedback in the colder climate, and a small positive lapse rate feedback due to low latitude ice sheets, however the combined feedback is close to that of $2xCO_2$. Source: Yoshimori et al. (2009). © American Meteorological Society. Used with permission

When the "'base climate" itself is very cold, (as in the LGM), perturbations *away from that climate* can show a weaker combined water vapour + lapse rate feedback than under the present climate. This results from weaker (less positive) high latitude feedbacks at high latitudes (Yoshimori, et al., 2011).

There is ample evidence that vapour feedback strengthens as the model base state warms from the current climate (Hu et al., 2017). Model experiments undertaken with "mixed layer oceans" (which equilibrate more rapidly than full ocean/atmosphere GCMs) find an increase of roughly 30% in feedback strength under forcing increasing from 2x to $16xCO_2$ (Meraner et al, 2013), at a rate across the range slightly higher than implied by fixed relative humidity (Colman and McAvaney, 2009). This increase has been shown to be due to a narrowing of the atmospheric window due to increased continuum absorption from water vapour (Seeley and Jeevanjee, 2021). At the same time (negative) lapse rate feedback strengthens too (Colman and McAvaney, 2009; Meraner et al., 2013), at least partially offsetting water vapour feedback increases (Jonko et al., 2013; Kluft et al., 2019). The lapse rate changes stem from strengthening negative tropical feedback from a continually steepening saturated adiabatic lapse rate and increased emission from upper tropospheric CO_2 (Seeley and Jeevanjee, 2021). The reduction then disappearance of high latitude positive feedbacks with accelerated loss of snow and sea ice cover with warming also strengthens the lapse rate feedback (Colman and McAvaney 2009). This offsetting has been attributed for models not projecting a "runaway" due to water vapour feedback even under extremely strong forcing, such as an increase in the solar constant of 25% (Boer, et al., 2005) or of CO_2 by a factor of 32 (Colman and McAvaney, 2009), as lapse rate feedback compensations result in a much more stable combined feedback (Colman and McAvaney 2009).

The processes behind strengthening water vapour feedback with temperature are now better understood. Using a combination of GCMs and a 1D RCM, Meraner et al. (2013) found the increase in tropopause height with temperature is critical, a finding similar to Rugenstein et al. (2019), although close to unchanged relative humidity remains an important process. Given the greater appreciation of the importance of processes around the tropopause, there may now be some caveats on earlier results (e.g. Boer, et al., 2005; Colman and McAvaney 2009) from model experiments with relatively coarse vertical resolution (Meraner et al., 2013).

The increased understanding of the spectral dependence of OLR with increasing temperature (see Section IV-C) also casts light on critical processes as surface temperature increases. Increasing CO_2 at very high levels of warming can dominate spectral cooling windows, thereby coupling OLR to tropospheric temperatures helping to stabilize global temperatures (Seeley and Jeevanjee, 2021). Note also that parametrised radiation schemes can become insufficiently accurate to properly resolve these processes as surface temperatures exceed around 310K (Kluft et al., 2021).

Water vapour and lapse rate feedbacks also show sensitivity to forcing *type*. In part this can originate from differences in horizontal and vertical forcing distributions but can also be affected by differences in absorption spectra, for example from less overlap of the O_3 absorption spectrum with water vapour compared to that of CO_2 (Yoshimori and Broccoli, 2008). Volcanic ejecta and sulphate forcing produce slightly weakened lapse rate and water vapour feedbacks, but again with the compensation producing close to unchanged combined feedback (Yoshimori and Broccoli, 2008). Other forcing agents including black carbon and tropospheric O_3 forcing strengthen (negative) lapse rate feedback compared with CO_2 forcing, although with substantial compensations from water vapour feedback changes (Rieger et al., 2017). Water vapour feedback is stronger for globally equivalent solar forcing compared to CO_2, as it is more strongly weighted to lower latitudes where feedback is strong, again though with some compensation from increased lapse rate feedback (Yoshimori and Broccoli, 2008). Although these studies used the traditional feedback decomposition, the compensating effects or water vapour and lapse rate feedbacks suggest the alternative RH based formulation would show little change in the Planck and lapse rate terms from the different forcings.

Low latitude lapse rate changes can be sensitive to the details of forcing, with evidence that changes in recent decades have been affected by the pattern of the anthropogenic and volcanic aerosol forcing (Santer et al., 2017). Modelling experiments find that lapse rate feedback is stronger (negative) under O_3 forcing then under the equivalent CO_2 (Rieger et al., 2017).

At *high latitudes*, different lapse rate responses may be induced by different radiative forcing, as for example found from CO_2 increases paired with reduced insolation from "solar radiation management" GCM results from GEOMIP, the Geoengineering Model Intercomparison Project (Robock et al., 2011). The CO_2 forcing produces a "bottom heavy" warming which outweighed lapse rate response to the more uniform solar forced change, or to advective changes (Henry and Merlis (2020). High latitude feedbacks are particularly complex and will be discussed at length in Section IV-I below.

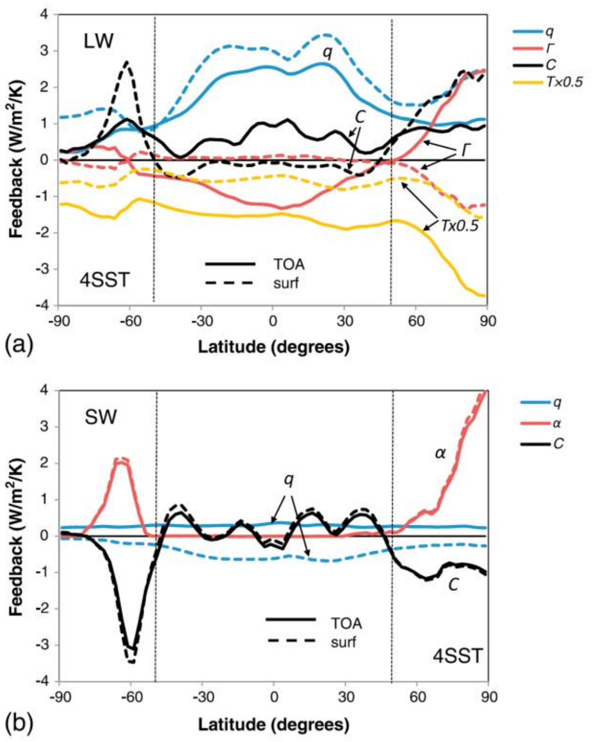


Figure 16. (Colour online) Latitude mean, annual radiative feedbacks defined at the TOA (solid lines) and at the surface (dashed lines) calculated for a GCM forced by SSTs from a equilibrium warming $4xCO_2$ forcing experiment ("4SST") for the (a) LW and (b) SW . Notation for feedbacks is as in Eqn 4, except "*T*" denotes the Planck feedback, *P*, and is scaled by a factor of 0.5 for display purposes.

This shows that at the surface (compared with TOA) SW water vapour feedback is reversed in sign, lapse rate feedback close to zero except at high latitudes, and water vapour feedback consistently stronger. Source: Colman (2015). Reproduced with permission from the American Geophysical Union.

H. A surface perspective on feedbacks

The conventional view of feedbacks is considered at the TOA, as this is fundamental to long-term planetary energy balance (Manabe and Strickler, 1964; Manabe and Wetherald, 1967) (see Section II). It is instructive, however, to also consider feedbacks at the Earth's surface as these provide different perspective and physical insights, and clarify the impact of feedbacks on features such as rainfall change under climate warming. The different components are listed in Eqns 9 and 10.

Zonally averaged surface and TOA feedbacks from one GCM are shown in Fig. 16 (Colman, 2015). In the LW, surface water vapour feedback is around 30-50% stronger than at the TOA (Pendergrass and Hartmann, 2014; Colman, 2015), which essentially renders the surface in radiative "runaway greenhouse" conditions as it exceeds the net cooling from the combination of surface blackbody cooling plus atmospheric downward Planck warming. The strong surface radiative warming is offset mainly by increased evaporation, which is a key process driving global precipitation increases with temperature (Andrews et al., 2009; Previdi, 2010; Pendergrass and Hartmann, 2014). From an atmospheric energy balance perspective, differences between TOA and surface feedbacks result in a change in net radiative heating requiring latent and sensible heat changes (coupled directly with surface evaporative adjustments) to restore heat balance (Previdi and Liepert, 2012).

In contrast to TOA water vapour feedback which is dominated by changes in the mid to upper troposphere, the contributions to surface feedback are strongly peaked in the lowest parts of the atmosphere, with negligible contributions above 500hPa (Previdi, 2010; Colman, 2015; Pendergrass et al., 2018; Kramer et al., 2019; Dacie et al., 2019). This is a consequence of the high LW opacity of the lower atmosphere (e.g. Shine and Sinha, 1991). This also means that surface feedbacks are relatively insensitive to changes in factors such as convective parameterisation, upwelling circulations and ozone distribution (Dacie et al., 2019). As for the TOA, surface water vapour feedback in models scales closely with unchanged relative humidity under warming (Pendergrass and Hartmann, 2014).

Observations support a strong positive surface water vapour feedback. A global study based on NCEP/NCAR reanalysis with an offline radiation calculation found confirming evidence of positive water vapour feedback at the surface where temperatures exceeded approximately 2° C (Lindberg, 2003). Radiation trend measurements from the European Alpine Surface Radiation Budget (ASRB) ground stations network, combined with correlations between surface temperature and ERA- 40 integrated water vapor confirm GHG warming accompanied by strong water vapour feedback at the surface (Philipona et al., 2004, 2005). Radiosonde trends in lower tropospheric water from 1964-1990, combined with temperature changes and surface radiative transfer calculations, also suggest strong positive water vapour feedback over this period (Prata, 2008).

Except at high latitudes, classically defined lapse rate feedback is weak at the surface (Colman, 2015; Kramer et al., 2019; Fig. 16) as the opacity of the lower atmosphere prevents mid-upper tropospheric warming having a direct surface radiative impact. There is some evidence of indirect surface impacts though, as enhanced upper tropospheric warming can contribute to increases in moisture in the lower troposphere which affects the surface radiation balance (Xiang et al., 2014).

As result of weakness in lapse rate feedback, the offsetting water vapour/lapse rate relationship found at the TOA is absent at the surface, so both feedbacks contribute to inter-model spread in net surface radiative response to forcing (Kramer et al., 2019), and hence to impacts such as changes in precipitation. In the SW, water vapour feedback provides a surface *cooling*, due to increased atmospheric absorption, of a magnitude slightly stronger than of TOA warming (Colman, 2015; Fig. 16). Note that it remains unclear what different insights a RH based surface feedback analysis would provide, as this promising approach is yet to be explored.

I. The role of lapse rate and water vapour feedbacks in regional climate variability and change

1. Polar amplification of warming

Greater than global average warming at high latitudes, so-called "polar amplification" is a ubiquitous feature in GCMs (Holland and Bitz, 2003) and is also found in observations (IPCC, 2019) and paleo records (Masson-Delmotte et al., 2006). For example, in years 100-150 after a CO_2 quadrupling, Arctic (60° to 90° north) warming averages 11.2° C compared with 4.3° C for the tropics (Pithan and Mauritsen, 2014; Fig. 17), with winter Arctic warming roughly double that of summer. There is strong evidence that lapse rate feedback in particular plays a strong role in this amplification.

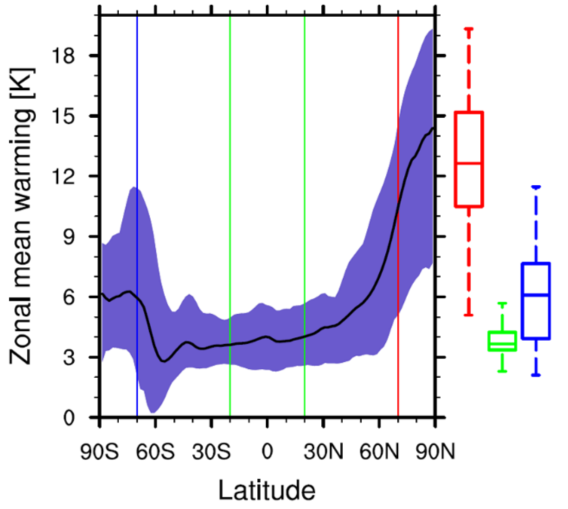


Figure 17. (Colour online) Surface temperature change from 14 CMIP5 models as a function of latitude, for years 100-150 after an abrupt 4xCO_2 forcing, showing the "amplification" in surface warming that occurs in the high latitudes, particularly in the Arctic. The thick black line is the model average, and the shading the full model range. Box-whisker plots left to right denote the warming averaged over , the Arctic poleward of 70 ° N, the Antarctic poleward of 70° S and 20° N-S. Box-whiskers represent 25th to 75th percentile and minimum, median and maximum values. Source: Block et al. (2020). Reproduced with permission, Tellus A, Taylor & Francis Ltd, www.tandfonline.com.

Polar amplification is of major consequence due to regional impacts from the accelerated warming (IPCC, 2019) and indeed has become "emblematic" of climate change (Boé et al., 2009). Moreover, the high latitude feedbacks and warming also have substantial global effects. Differences in strength between models in the surface albedo and lapse rate feedbacks at high latitudes in turn affect meridional temperature gradients and associated heat fluxes, thereby contributing to differences in low latitude circulation such as Hadley cell overturning (Feldl et al, 2017b). Furthermore, large volumes of consequential land ice melting of could lead to large sea level rise (IPCC, 2019).

The reasons for polar amplification are complex, involving a balance between changes in surface albedo, lapse rate, water vapour and cloud feedbacks, as well as in atmospheric and oceanic poleward fluxes, and these can vary widely between models (Ramanathan, 1977; Curry et al., 1995; Crook et al., 2011; Ghatak and Miller, 2013; Graversen et al., 2014a; Pithan and Mauritsen, 2014; Feldl et al., 2017a, b; Mokhov et al, 2016; Payne et al., 2015; Stueker et al., 2018; Block et al., 2020). Considering feedbacks both from a regional and hemispheric perspective provides valuable insights into processes important for poleward amplification and its variation across GCMs (Feldl and Roe, 2013).

From a *hemispheric and TOA perspective*, a key driver is changes in radiative imbalances induced by latitudinal variation in feedbacks, which in turn affect poleward atmospheric/oceanic heat fluxes (Zelinka and Hartmann, 2012). From this perspective lapse rate feedback has been shown to be the greatest contributor to annual mean polar amplification in CMIP5 models, as it cooled the tropics but warmed high latitudes (Pithan and Mauritsen, 2014; Steucker et al., 2018) – Figs. 9b, 18a.

The Planck feedback has also been hypothesised to contribute to amplification, because of the strong dependence of OLR increase per degree of warming in the warm tropics compared to the cold high latitudes (Pithan and Mauritsen, 2014) but this has been challenged by other studies that point to the importance of atmospheric emission temperatures rather than the surface, suggesting the Planck feedback gradient may even reduce polar amplification (Feldl and Roe, 2013; Henry and Meerlis, 2019). Considering feedbacks in the "fixed relative humidity" framework (Held and Shell, 2012; Eqns 6-8) retains the alternative "lapse rate" feedback as the most important contributor to polar amplification, but now the redefined Planck feedback contribution is small (Pithan and Mauritsen, 2014).

From the hemispheric TOA viewpoint, water vapour feedback, although warming the Arctic in absolute terms, opposes polar *amplification*, since it is much stronger at low latitudes than high (Langren et al., 2012; Zelinka and Hartmann, 2012; Taylor et al., 2013)..

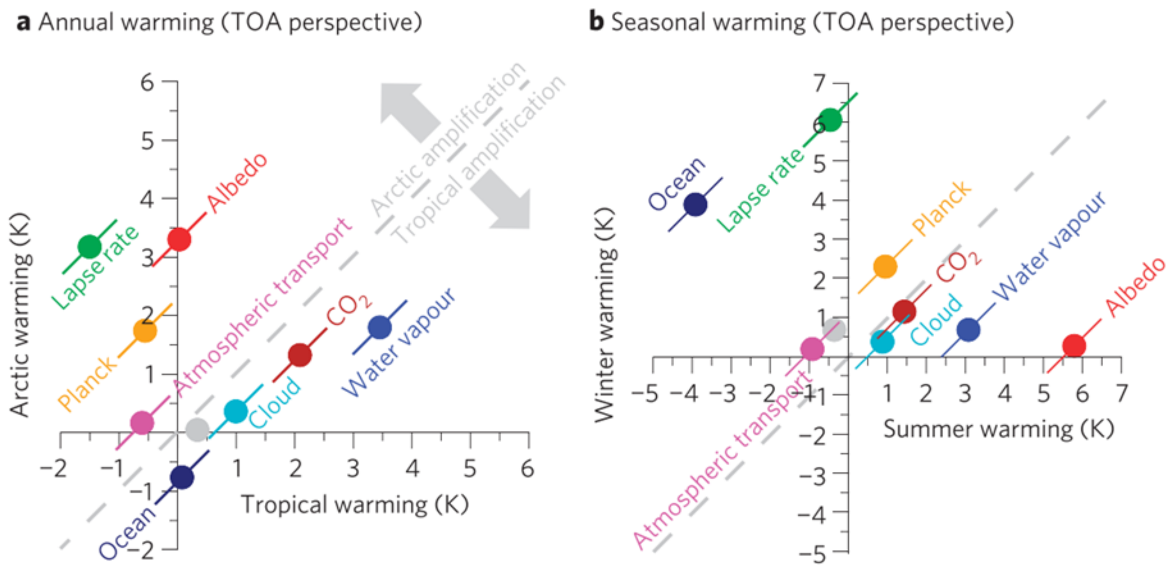


Figure 18. (Colour online) (a) Contribution to Arctic warming versus tropical warming in 16 CMIP5 models from diagnosed TOA feedbacks (lapse rate, water vapour, Planck, surface albedo and cloud) from changes in atmospheric transport and ocean uptake/transport, and from latitudinal dependence of CO_2 forcing. Values to the top left of the grey dashed line increase polar amplification, and to the bottom right decrease it. The importance of lapse rate feedback is apparent from its warming of the Arctic, versus tropical cooling, whereas water vapour feedback warms the tropics more. (b) seasonal variation shown by winter versus summer

warming from each of the processes in (a). The contrasting seasonal roles of lapse rate (winter) and water vapour (summer) are apparent. The grey dot represents the residual from total warming minus the addition of the individual components. Source: Pithan and Mauritsen (2014). Reprinted by permission from Springer Nature.

At high latitudes the presence of cold dense air near the surface (particularly during the cool seasons) can induce weak coupling between the surface and the free atmosphere. Lapse rate feedback is therefore a key contributor to (particularly winter) polar amplification because the highly stable lapse rate acts to trap warming in the lowest atmospheric levels, increasing surface warming relative to the layers above (Manabe and Wetherald, 1975). This vertical decoupling means that it is important to also consider *both a surface and a regional feedback view* in understanding feedback contribution to polar amplification (Taylor et al, 2013; Pithan and Mauritsen, 2014; Laîné et al., 2016). From this perspective, the largest Arctic warming results from greater downward than upward LW at the surface, again due to the nonlinear dependence on temperature of blackbody emissions (Pithan and Mauritsen, 2014; Sejas and Cai, 2016), and again from this perspective water vapour feedback reduces polar amplification.

Model experiments find that at a *regional level*, high latitude lapse rate and surface albedo feedbacks interact to amplify each other: strengthened surface albedo feedback results in a warmer surface and stronger positive lapse rate feedback, which can strengthen surface warming, further melting snow and sea ice (Döscher et al., 2014; Graversen et al., 2014a). Indeed, although the surface albedo feedback is important, Arctic amplification can occur without it (Hall, 2004; Graversen and Wang, 2009; Kim et al., 2018; Russotto and Biasutti, 2020), and LW feedbacks are known to play a dominant role in the region in coupled models (Winton, 2006). Suppression of lapse rate feedback (by locking lapse rates) in a GCM reduced Arctic amplification of warming by 15%, and Antarctic by 20%, although interaction with surface albedo feedback meant that it could not be properly considered as a separate feedback process (Graversen et al., 2014b).

The role of *heat transport* by atmosphere and ocean, and its interaction with feedbacks, has also been intensively investigated. Although poleward heat transports are important in maintaining energy balance, and contribute to warming, model *spread* in polar amplification is primarily due to differences in feedbacks (Hwang et al., 2011; Stuecker et al, 2018). Atmospheric heat flux changes in fact act to reduce model spread by opposing radiatively induced differences, and ocean heat transport changes are uncorrelated with warming across models (Pithan and Mauritsen, 2014).

Observations of feedbacks poleward of 60° N for the period 2000 to 2014 indeed find a dominant role for lapse rate feedback in the positive LW feedback, with little contribution from water vapour (Hwang et al., 2018, see Fig. 19). Hwang et al. (2018) explain the weakness of the water vapour feedback as due to moistening being confined to the lower parts of the atmosphere, whereas lapse rate feedback is strong due to temperature differentials between the surface and atmospheric upper levels. Observations also suggest that on *short timescales* (monthly to interannual), lapse rate and surface albedo feedbacks are of comparable magnitude, but that lapse rate feedback makes the greatest contribution to high latitude amplification, and water vapour feedback opposes it (Zhang et al., 2018). The strength of lapse rate feedback, along with other feedbacks can also be expected to further change under ongoing loss of Arctic sea ice (Dekker et al., 2019).

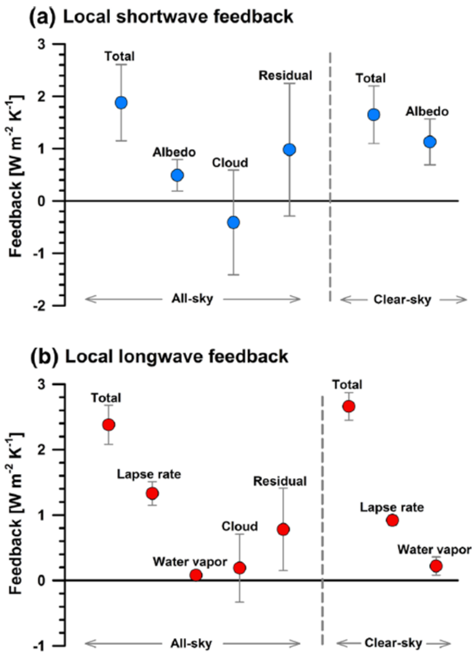


Figure 19. (Colour online) Radiative feedbacks diagnosed using 2000-2014 Clouds and the Earth's Radiant Energy System (CERES, Wielicki et al., 1996) TOA SW and LW fluxes combined with surface temperature and vertical profiles of temperature and humidity taken from ERA-Interim reanalyses (Dee et al., 2011). Values shown are calculated by linear regression of TOA fluxes with regional surface temperature and kernels applied to temperature and moisture profiles to diagnose individual feedbacks. Error bars show standard error from regressions combined with estimated CERES uncertainties. "Clear sky" represents results with all cloud effects removed. The importance of lapse rate feedback is apparent in regional warming, whereas water vapour contributions are diagnosed as small. From: Hwang et al. (2018). Reprinted by permission from Springer Nature.

Model studies find that the magnitude of the polar amplification and the role of different feedbacks change throughout the year (Block et al., 2020). Polar lapse rate feedback is positive and reinforcing of amplification in winter to spring, when the atmosphere is dominated by inversions, but is negative and weakens amplification in summer to autumn, when surface inversions are weaker or absent (Colman, 2003b; Pithan and Mauritsen, 2014; Fig. 18b; Kim et al., 2018). Differences in climatological polar cloud fractions, for example from cloud parameterisation changes, can also strongly affect the seasonality of both lapse rate and water vapour feedbacks (Kim et al., 2016).

The strength of the high latitude lapse rate feedback depends not just on season, however, but on the *magnitude* of the high latitude warming (Feldl et al., 2017a). If warming is strong, a more positive lapse rate feedback further enhances surface temperature increase and polar amplification. If surface warming is only moderate, then flux convergence from atmospheric eddies contribute to regional stabilisation and neutral (or even negative) high latitude lapse rate feedback.

The polar lapse rate feedback contribution to warming also depends on the *nature and profile* of the forcing, as it is not uniquely dependent of the surface temperature, but also on other regional processes (Cronin and Jansen, 2016; Henry and Merlis, 2020). Lapse rate response differs between surface forcing, such as from CO_2 changes, a reduction in surface albedo or increase in oceanic heat transport compared with forcing such as from increased poleward advection or increased atmospheric SW absorption (Cronin and Jansen, 2016). This implies that advective heat fluxes from lower latitudes can also play an important part in regional heat balance adjustment. If additional atmospheric heat flux convergence from lower latitudes "wins out" over SW surface warming from reduced albedo, this can result in surface stabilisation and a negative lapse rate feedback (Cronin and Jansen, 2016). Consistent with this, investigation of the impact of CO_2 forcing, offset by "geoengineered" SW flux reductions found positive lapse rate feedback associated with the "bottom heavy" warming from the CO_2 response won out over the atmospheric SW induced changes, meaning polar amplification persisted despite the global radiative balance (Henry and Merlis, 2020)

Despite the major role of lapse rate feedback in causing the polar amplification in models, when looking across models the *spread* in net radiative feedback in the Arctic arises more from differences in Planck and albedo feedbacks rather than lapse rate (Block et al., 2020). Other feedbacks may also play a role. The cloud feedback impact on polar amplification appears modest (Pithan and Mauritsen, 2014, Fig. 18), but clouds interact to strengthen or weaken lapse rate feedback from changes in downward LW radiation (Tan and Storelvmo, 2019). The role of the stratosphere in polar amplification has been relatively little studied. A recent GCM experiment found that stratospheric water vapour feedback was three times stronger at high latitudes than low, contributing around 14% to Arctic amplification (Li and Newman, 2020).

Models consistently project a greater amplification in the Arctic than the Antarctic under transient climate change (Manabe et al., 1991; Fig. 17), in part due to of deep mixing in the Southern Ocean which slows the warming (IPCC, 2019). Experiments both with and without Antarctic ice sheet elevation indicate comparable *equilibrium* amplification to the Arctic with no Antarctic topography (Salzmann, 2017; Hahn et al., 2020), with the implication that smaller equilibrium amplification in the Antarctic results from a weaker, shallower temperature inversion related to topographic damping of meridional heat fluxes and cooling from katabatic winds (Hahn et al., 2020). Consequently, lapse rate feedback pays a smaller role in seasonal variation in Antarctic temperature amplification (Hahn et al., 2020).

In summary, observations and a large number of modelling studies confirm that lapse rate feedback is a critical factor in polar amplification both for its global-scale TOA structure (negative feedback at low latitudes, positive at high), and also for its regional scale surface impacts and interactions. Water vapour feedback although important for broadscale warming generally opposes the amplification, largely due to its greater strength at low latitudes than high. A large number of methodological approaches and differences in models and datasets produces somewhat different quantifications of these processes, however, and preclude a simple unifying conceptual framework and unambiguous quantification of feedback impact. Further refining this understanding and framework remains as an ongoing challenge (Russotto and Biasutti (2020).

2. Climate features and regional variability

Apart from global or broad scale radiation changes, such as over high latitudes, there is ample evidence that water vapour and/or lapse rate feedback can play important roles in the characteristics of "modes of variability" (i.e. of preferred patterns of large scale spatio-temporal variability).

One example is the El Niño Southern Oscillation (ENSO), where the presence of water vapour/LW interaction affects the vertical structure of radiative heating associated with surface temperature anomalies in the tropics, and acts to amplify ENSO variability (Hall and Manabe, 2000b).

The seasonal movement of the Inter-Tropical Convergence Zone[14], (ITCZ), provides a second example in which water vapour feedback plays a role. Water vapour feedback processes have been found to roughly double the seasonal movement of the inter-tropical convergence zone, ITCZ (Clark et al., 2018), which was traced to changes the inter-hemispheric asymmetry of subtropical relative humidity (Peterson and Boos, 2020). Water vapour and lapse rate feedbacks also play a role in maintaining the Hadley circulation response to asymmetric forcing, such as hemispheric warming (Yoshimori and Broccoli, 2009). Aqua planet experiments using "radiation suppression" found that under CO_2 forcing, water vapour feedback widens the monsoon region, and increases monsoon associated moisture and rainfall, by warming and moistening the region (Byrne and Zanna, 2020). There is evidence, too, that regional water vapour feedback plays an important role in persisting anomalously high winter SSTs into subsequent seasons in the tropical North Atlantic region mainly responsible for the genesis of Hurricanes (Wang et al., 2017), on top of well-known cloud-SST and wind-evaporation-SST feedbacks.

Moisture-radiative feedbacks also play an important role in the dynamics of some tropical intra-seasonal processes (Bony and Emanuel, 2005). Regional water vapour and lapse rate feedbacks are important in the propagation and magnitude of the Madden-Julian Oscillation (MJO)

[14] The region of enhanced convection, lying close to the equator, representing the upward branch of the Hadley Circulation.

– a near equatorial ~30-60 day wave featuring coupling between convective and circulation processes (Madden and Julian, 1994; Hendon and Salby, 1994). This includes enabling the MJO to penetrate further into the Maritime Continent (Indonesian region) "barrier" due to stronger heating resulting from water vapour feedback coincident with the convective envelope (Zhang et al., 2019).

There is some evidence that regional water vapour feedback may also amplify regional surface responses to warming. Elevation dependent water vapour feedback has been proposed as being partly responsible (along with surface albedo feedback and other processes) for observed amplification of climate change warming with altitude in mountainous regions (Pepin et al., 2015). The hypothesised physical process is that due to decreasing water vapour with altitude, absorption bands are under saturated (because of less overlying total water vapour), resulting in larger additional down-welling LW radiation under surface warming (Rangwala et al., 2009, 2010; Rangwala, 2013; Rangwala et al., 2013; Palazzi et al., 2017). The feedback loop is evidenced by statistical relationships between water vapour, down-welling LW radiation and surface warming (e.g. Rangwala et al., 2009). These findings must be treated with some caution, however, as a recent high-resolution study of warming in the Rocky Mountains, using a Regional Climate Model (RCM) found no evidence of amplification by elevation dependent water vapour feedback (Minder et al., 2018).

Similarly, local surface water vapour feedback has been hypothesised to play a role in regional temperature variability, such as in response to ENSO (Zhang et al., 2011) and temperature extremes such as heatwaves (Oueslati et al., 2017). There is some evidence too that despite low humidities, surface water vapour feedback enhances the dryness of desert regions from strong coupling between the surface and lower atmosphere, and the particular sensitivity of downward LW radiation to water vapour increases in very dry environments (Zhou, 2016; Zhou et al., 2016; Wei et al., 2017).

It is not clear to what extent these latter processes are fully established, or indeed truly "closed loop" feedbacks, rather than water vapour responses to/drivers of large-scale forcing and variability, and they are not discussed further here. Further research would be needed to fully establish their veracity and importance.

V. Observational evidence for water vapour and lapse rate feedback

A. Moisture trends and variability in the lower atmosphere

As global temperature has risen by around 0.5° C since 1990 (Masson-Delmotte et al., 2018), we might expect to see changes in global water vapour amount and distribution under that warming.

Water vapour changes in the lower atmosphere are strongly coupled with the surface (Trenberth et al., 2005), and expected strong increases with temperature have been confirmed by observations. Globally, satellite and radiosonde analyses confirm (lower tropospheric dominated) "total precipitable water[15]" variations consistent with close to unchanged relative humidity under interannual variability (Wentz and Schabel, 2000; Dai et al., 2010; Trenberth et al., 2015).

Over oceans, energy balance arguments suggest only small changes of relative humidity would be expected with increased temperature (Jeevanjee, 2018). This is because significant shifts in relative humidity would imply large changes in evaporation (in the absences of large changes in wind or stability), which cannot be sustained energetically, as the subsequent latent heat release in the troposphere cannot be matched by radiative cooling (Allen and Ingram, 2002; Held and Soden, 2014). Consistent with this, models project only small *trends* in lower tropospheric relative humidity with secular warming, that of modest increases (Byrne and O'Gorman, 2013). Observations provide strong

[15] Vertically integrated water vapour content

evidence for increases in total precipitable water over ocean regions from satellite-based Special Sensor Microwave Imager (SSM/I) data (Santer et al., 2007; Wentz et al., 2007). Trends in *relative humidity* over oceans are less clear (Willett et al, 2008), but appear broadly consistent with unchanged relative humidity in line with underlying SSTs (Byrne and O'Gorman, 2013; Hartmann et al, 2013). Two hourly Global Positioning Gystem (GPS, see Section V-B) measurements show an increase in precipitable water from 1995 to 2011, also roughly in line with unchanging relative humidity, along with change that is larger at night than during the daytime (Wang et al, 2016).

Over most land areas, models predict decreasing relative humidity in the lowest parts of the atmosphere, particularly during the warm season (O'Gorman and Muller, 2010; Byrne and O'Gorman, 2016). Observations broadly confirm this trend, except for some regions in the tropics and high northern latitudes (Willett et al., 2014).

B. Upper tropospheric moisture

Although these robust responses in the lower troposphere provide important confirmation of increasing specific humidity with a warming climate, it is *mid to upper tropospheric humidity* that is most important for water vapour feedback (Section IV-D) and water vapour trends in this region are less straightforward to measure. Over recent decades several different observational approaches have been taken to monitor variability and change in this challenging region. Confidence in the results depends upon the robustness of the measurement methodology, so they are briefly reviewed here.

1. Methods of measuring upper tropospheric humidity.

There are three principal observational sources for monitoring trends and variability of upper tropospheric humidity: the radiosonde network, satellite measurements and atmospheric "reanalyses".

The radiosonde network of balloon-borne soundings has been long established to provide vertical profiles of temperature and moisture for input into operational numerical weather prediction systems. In principle, radiosondes can measure changes in humidity at a much finer vertical resolution than satellites. However, attempts to use the radiosonde network for long-term climate monitoring and detection purposes have encountered several major challenges.

The first has been moisture biases, including temporally and spatially varying biases from instrumental and measurement technique differences between countries and changes over time (Parker and Cox, 1995; Seidel et al., 2009). Accuracy problems are widespread. For example, even to recent decades dry biases of up to 20% have been evident in the middle troposphere from commonly used radiosondes (Miloshevich et al., 2009), and biases can also result from the emergence of radiosondes from saturated regions into much drier overlying layers (Held and Soden, 2000). A further issue is the inherent data sparse nature of the radiosonde network leaving large tropical and oceanic regions, for example, severely under sampled (Müller et al., 2016), with sampling particularly limited in the stratosphere (Hurst et al., 2011; Hegglin et al., 2014).

To address these issues there have been several separate efforts to homogenise global operational radiosonde observations (e.g. Durre et al., 2009, McCarthy et al., 2009; Dai et al., 2011). Indeed, the limitations inherent with the radiosonde network, together with the recognised importance of monitoring upper tropospheric and lower stratospheric water vapour have led to calls for the development of global, carefully calibrated and long-term balloon-born upper troposphere water measurement program (Müller et al., 2016).

Satellite measurements have been increasingly used over recent decades to estimate variability and trends in relative humidity.

Upper troposphere humidity can be inferred from instrument such as the 6.7μm radiance channel from the High-Resolution Infrared Radiation Sounder (HIRS), which is sensitive to moisture in a deep upper troposphere layer from roughly 200-500 hPa (Soden et al., 2000). HIRS measurements have been made for over 40 years since the launch of the Television Infrared Observation Satellite (TIROS-N) in 1978 (Shi and Bates, 2011). Unfortunately, a break in the TIROS-N record occurs in 2005, when the central wavelength of the HIRS instrument was changed from 6.7 μm to 6.5 μm, limiting the record to 27 years (1979-2005) (Chung et al., 2014), although comparable measurements have been available from the microwave sounder SAPHIR from 2011 (Brogniez et al., 2015). Since the purpose of the HIRS mission was weather prediction, not climate monitoring, producing very long, multi-satellite trends is challenging because of inter-satellite biases (John et al., 2011), and careful bias correction has been needed to produce a continuous, consistent dataset suited to climate applications (Bates and Jackson, 2001; Jackson and Soden, 2007; Shi and Bates, 2011). The difference between MSU/AMSU channel 2 brightness temperatures and HIRS channel 12 is also useful for removing defective temperature changes on the upper troposphere to produce a cleaner measure of upper troposphere relative humidity (Chung et al., 2014).

Other important satellite datasets derive from the Microwave Limb Sounder (MLS), and the newer Cross-track Infrared Sounder. These instruments provide high-quality temperature and water vapour profiles from as early as 2002 and these measurements have been used extensively to study water vapour variability and trends (e.g., Dessler and Minschwaner, 2007; Liu et al., 2018).

Global Positioning System networks can also be exploited to produce water vapour datasets, essentially measuring the integrated temperature and humidity along the GPS path length (Jin et al., 2007; Wang and Zhang, 2008; Wang et al., 2007; Wang et al., 2016; Vergados et al., 2016). This is a relatively new and short dataset, with around 100 ground stations established in 1997. The GPS technique has inherent advantages, including not requiring calibration, and being essentially unaffected by clouds (Sherwood et al., 2010b). Trends derived from GPS data indicate moistening Ho et al., 2018), however these can be sensitive to beginning and end values (i.e. variability) given the shortness of the GPS timeseries (Hartmann et al., 2013).

Atmospheric reanalyses provide a source of what might be considered "pseudo-observations". These are produced by running recent-version numerical weather prediction models on observations from past years/decades retrieved from extensive archived data sources. By exploiting advanced data assimilation techniques, they produce a climate as closely as possible constrained by those observations (Slingo et al., 1998). As such, they represent a "fixed model" representation of past climate, but remain subject to inherent model deficiencies, particularly in data sparse areas, and are subject to a greatly varying input dataset in terms of observational instrumentation, coverage and accuracy (Thorne and Vose, 2010; Fujiwara et al., 2017). Because of these changes in the observational network, or because of limitations on data ingestion or data quality trends must be treated with some caution (Dessler et al., 2008; Dessler and Davis, 2010). Nevertheless, they produce a convenient and comprehensive, observationally constrained and physically consistent estimate of past climate gleaned from a vast store of observational datasets.

2. Trends in upper tropospheric humidity

A range of studies have concluded that long-term trends are consistent with near unchanged relative humidity in the upper troposphere (Allan et al., 2003; Soden et al., 2005; Cess, 2005; Ferraro et al., 2015). After careful homogenisation and bias correction of the HIRS brightness temperatures from both TIROS-N and Metop-A satellites, Shi and Bates (2011) found little change in equatorial tropical upper tropospheric relative humidity over the 30-year period from 1979 to 2008. On top of such broadscale trends, of course lie superimposed regional and latitudinal changes due to changes in circulations (Bates and Jackson, 2001). Allowing for these is important in understanding long term broad-scale water vapour feedback and extreme caution must be exercised when considering limited-region data and extrapolating to global means (Dai et al., 2011).

An examination of four more recent reanalysis products (ERA40, Japanese Reanalysis (JRA), Modern Era Retrospective- Analysis for Research and Applications (MERRA, Rienecker et al., 2011), and the European Centre for Medium- Range Weather Forecasts (ECMWF)- interim reanalyses, Dee et al., 2011) found unanimous agreement on increasing humidity between 1984 and 2009 as well as with ENSO-induced warm interannual fluctuations (Dessler et al., 2008; Dessler and Davis, 2010).

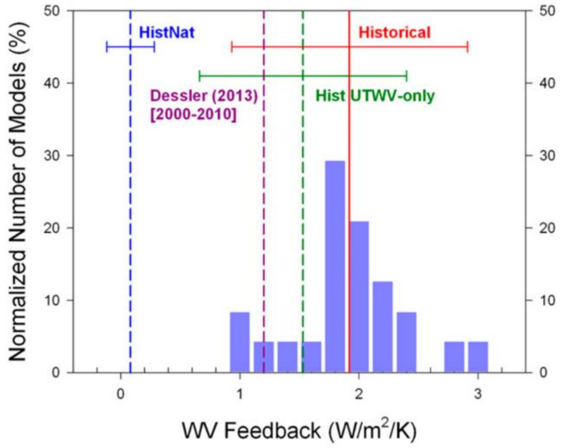


Figure 20. (Colour online) Attributing observed water vapour feedback to anthropogenic greenhouse gas emissions: Bars show water vapour feedback between the periods 1979-1988 and 1989-1998 from CMIP5 experiments forced by: natural forcing agents due to solar and volcanic changes only ("HistNat"); anthropogenic greenhouse gases and aerosols, as well as natural forcings ("Historical"). A third calculation is performed ("Hist UTWV only"), derived from the historical runs but using only humidity changes occurring above 600 hPa.

Differences between "Historical" and "HistNat" show that models do not match observational estimates of water vapour feedback *unless* anthropogenic greenhouse gas forcing is included: i.e. this feedback can be "attributed to" anthropogenic emissions. Differences between "Historical" and "Hist UTWV only" demonstrate the dominance of the upper troposphere, in that around 80% of the feedback strength can be attributed to upper humidity changes in this region.

One estimate of water vapour feedback from reanalyses (Dessler, 2013) is marked in purple. See Tables 1 and A2-1 for other estimates of water vapour feedback from unforced variability and climate change. Source: Chung et al. (2014).

But is the moistening due to human activities? An "attribution" study by Chung et al. (2014) considered satellite derived tropical upper tropospheric humidity trends over 27 years (to 2005) with CMIP5 model simulations (Fig. 1920). They concluded that observed changes were only consistent with the models in which the forcing applied included anthropogenic GHGs—i.e. changes were absent when models saw "natural" forcing by volcanic sulphates and solar changes alone. Historical feedbacks are also consistent with feedbacks calculated assuming fixed upper tropospheric relative humidity (Fig. 20) (Chung et al., 2014). This result is important because it provides strong evidence not just that water vapour feedback is occurring and is strongly related to upper tropospheric humidity trends, *but that it is operating in response to human-induced warming.*

C. Use of variability analogues for evaluating water vapour feedback.

In addition to measurements of humidity *trends*, many observational studies have been made for the relative humidity response under "natural" (i.e. unforced) *variability* including from the seasonal cycle and interannual and decadal variability. This has the advantage of providing observable tests for water vapour response to temperature change and avoids possible pitfalls in deriving reliable, long-term homogeneous data series and detecting modest trends. Caution is needed however, as discussed below, in interpreting the resultant feedback as analogues for water vapour feedback seen under long term climate change, largely because of differences in SST patterns associated with global temperature changes.

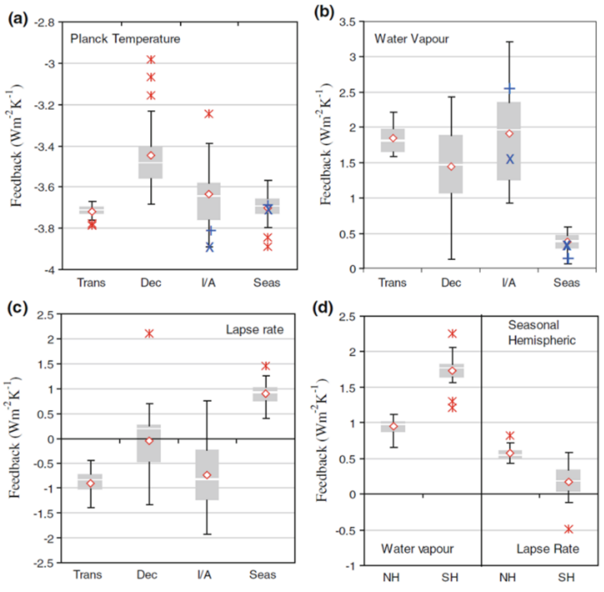


Figure 21. (Colour online) Planck, water vapour and lapse rate feedbacks from CMIP5 models, calculated using PRP for: "transient" (i.e. secular) response to increased CO_2 ("Trans") and from unforced decadal ("Dec"), interannual ("I/A") and seasonal ("Seas") variability. In panel (d) the seasonal feedback is broken up into separate northern and southern hemispheres. Box whisker plots show median, 25-75th percentiles, ranges within 1.5 inter-quartiles, and outliers (stars). X's (+'s) show results calculated from ERA40 (MERRA) reanalyses. From: Colman and Hanson (2013). Reprinted by permission from Springer Nature.

On the **seasonal cycle**, large hemispheric scale changes in temperature occur, with observed winter-summer water vapour changes consistent with close-to-unchanged relative humidity, including in the upper troposphere (Rind et al., 1991). Satellite derived OLR changes show radiative damping (below Planck cooling) from strong positive water vapour feedback (Tsushima et al., 2005) and water vapour feedback from ERA40 and MERRA reanalyses are of comparable strength to that of CMIP5 GCMs, both globally and on hemispheric scales (Colman and Hanson, 2013; Fig. 21d). These results are all consistent with a feedback from unchanged relative humidity. A caveat, however, is that the large hemispheric temperatures swings do not bear close resemblance to patterns of changes under global warming, and the large, compensating positive and negative radiative responses in the warmer and cooler hemispheres result in a relatively weak global net feedback (Colman and Hanson, 2013; Fig. 21b).

On **interannual timescales** too, caution must be exercised as the pattern of warming associated with global temperature change is different under climate change warming and unforced variability. Regional relative humidity fluctuations and associated TOA radiative changes in the tropics follow large-scale interannual circulation changes, such as those associated with ENSO (Blankenship and Wilheit, 2001; Bates et al., 2001; Brown et al., 2016; Tian et al., 2019) or planetary scale mid-latitude atmospheric waves (Bates and Jackson, 2001). Consequently, interannual surface temperature fluctuations are much more strongly peaked close to the equator, than are temperature increases under global warming, consistent with ENSO-related SST changes (Hurley and Galewsky, 2010; Colman and Hanson, 2013; Dessler, 2013). These result in a strong low latitude peak in the LW water vapour radiative response (Dessler and Wong, 2009), as shown in Fig. 22.

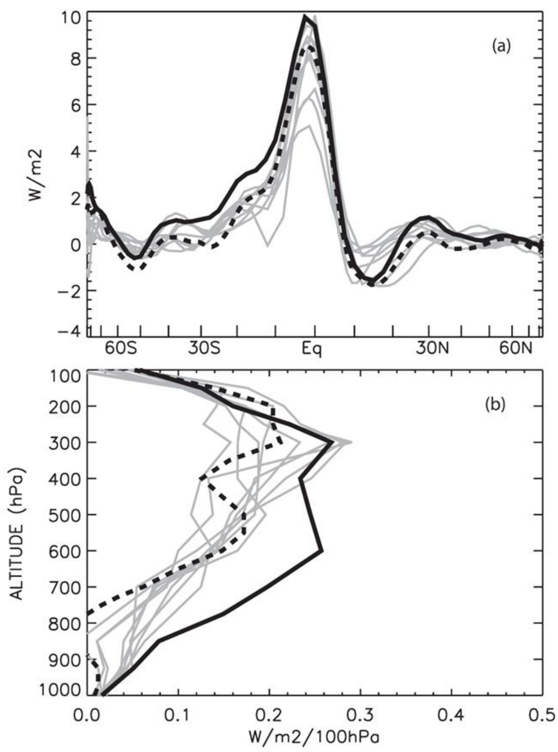


Figure 22: (a) longitudinally averaged TOA flux responses from water vapour changes over an ENSO cycle as depicted by 12 CMIP3 models

(grey lines) and derived from two reanalyses – MERRA (dashed black) and ERA40 (solid black). (b) flux changes (in $Wm^{-2}100hPa^{-1}$) from water vapour perturbations as a function of altitude. Source: Dessler and Wong (2009).

Since much of the variability in upper tropospheric humidity is driven by ENSO-related migration of convective features such as the Inter-Tropical Convergence Zone (ITCZ) (Xavier et al., 2010), and modified regionally by deep convection (Su et al., 2006), it is important to consider changes over very large areas, which include both rising and descending regions. When calculated over such large spatial scales, the data are consistent with close to constant relative humidity (Gettelman and Fu, 2008; Dessler et al., 2008; Chung et al., 2010). Averaging over the entire tropics, 6.7 μm HIRS brightness temperature responses are also consistent with close to unchanged relative humidity (Allan et al., 2003; McCarthy and Toumi, 2004), although another study using the HALOE MLS found modest decreases in tropics averaged relative humidity associated with temperature increases in convective regions (Minschwaner and Dessler, 2004). A recent comparison of tropical mean 200hPa specific humidity variations with temperature as measured from three datasets: GPS refractive indices, AIRS satellite retrievals and the MERRA reanalysis also found values consistent with small reductions in relative humidity (Vergados et al., 2016; Fig. 23). Together these studies make an overwhelming case of broadscale upper tropospheric humidity responding to global surface temperature perturbations close to, or slightly, below fixed relative humidity values.

Because of the more peaked tropical warming, radiative response under interannual variability may be expected to exaggerate the strength of the resultant feedback compared with climate change feedback (Dessler 2014; Colman and Hanson, 2016; Po-Chedley et al., 2018; contrast Figs 17a and 22a). In fact, results show roughly comparable strength seen in water vapour feedback from interannual and secular climate change, albeit with some showing interannual somewhat stronger (Colman and Hanson, 2013; Fig. 21), while others a little weaker (Koumoutsaris, 2013; Liu et al., 2018). Much more limited literature suggests decadal water vapour feedback slightly weaker than for interannual (Colman and Hanson, 2013; see Fig. 21).

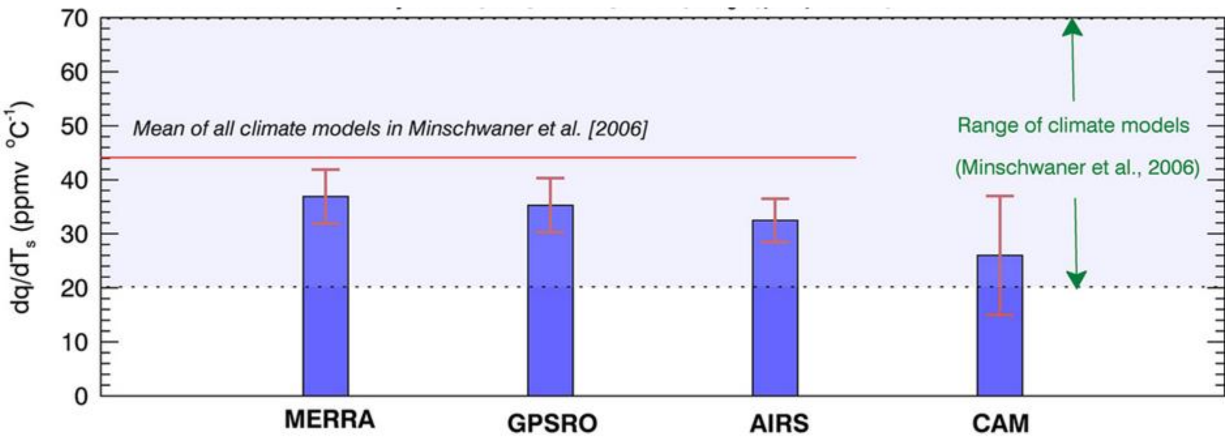


Figure 23. (Colour online) Estimated specific humidity variations with temperature averaged over the tropics from MERRA (Rienecker et al., 2011), Atmospheric Infrared Sounder (AIRS) satellite retrievals, and 1.2-1.6 GHz Global Positioning System Radio Occultation (GPSRO) measurements at 250 hPa over the period 2007-2010. This shows close agreement between the three observational measurements. The values of dq/dT_s indicate moistening with temperature at slightly below the rate implied by unchanged relative humidity. Error bars represent one standard deviation estimation uncertainty from linear regressions. Also shown are calculations from a single GCM, the Community Atmosphere Model, CAM (Gettleman and Fu, 2008).

The solid line and grey area represent the mean and spread of dq/dT_s in 42 CMIP5 GCMs from Minschwaner et al. (2006), and show overall consistency between models and observations, albeit with wide model spread, and mean model value a little greater than observational estimates. Source: Vergados et al. (2016). Reproduced with permission from the American Geophysical Union.

Another possible difference is that the contribution to the interannual TOA radiative response may be less strongly peaked in the upper troposphere than for climate change (Hall and Manabe, 1999; Dessler and Wong, 2009; Colman and Power, 2010, see Fig. 22). However, some studies find strong upper tropospheric peaking for interannual feedback (Colman and Hanson, 2013) and reanalyses derived feedbacks can diverge strongly (Dessler and Wong, 2009; Dessler 2010; Colman and Hanson, 2013; Fig. 22), so this possible structural difference remains unclear.

Despite the differences in geographical and (perhaps) vertical structure, an important finding has been that a modest correlation exists between individual CMIP5 model climate change and interannual LW water vapour feedbacks (Gordon et al., 2013; Colman and Hanson, 2016; Takahashi et al., 2016; Liu et al., 2018; Dalton and Shell, 2013). This provides strong motivation to estimate water vapour feedback from observations as it may allow a semi-direct evaluation of climate change feedback on top of providing a key test of water vapour processes in models via the strength and structure of their interannual feedbacks.

A significant number of estimates of the strength of interannual water vapour feedback have now been made, as summarised in Table 1. Tellingly, published estimates show substantial divergence in diagnosed feedback magnitude—although importantly all agree on a strong positive feedback on interannual timescales.

Reference	Dataset(s)	Analysis period	Value ($Wm^{-2}K^{-1}$)
Water vapour			
Dessler and Wong (2009)	ERA-40 (Uppala et al., 2005 MERRA	ENSO warm/cold phases, 1980-2000	3.7 (Net) 4.7 (Net)
Dessler (2013)	ERA Interim (Dee et al., 2011) MERRA (Rienecker et al., 2011)	2000-2010	1.35 (Net) 1.12 (Net)
Colman and Hanson (2012)	ERA Interim MERRA	1960-1998 1980-2008	1.6 (LW) 2.5 (LW)
Gordon et al. (2013)	Atmospheric Infrared Sounder AIRS-MLS	2002-2009	2.2 ± 0.4 (Net)
Koumoutsaris, (2013)	JRA-25 (Onogi et al., 2007) ERA-Interim	1979-2009 1979-2009	0.86 ± 0.14 (Net) 1.37 ± 0.16 (Net)
Liu et al. (2018)	AIRS-MLS	2004-2016	1.46 ± 0.22 (LW) 0.09 ± 0.01 (SW) 1.55 ± 0.23 (Net)
Lapse rate			
Koumoutsaris, (2013)	JRA-25 ERA-Interim	1979-2009 1979-2009	0.34 ± 0.20 0.11 ± 0.16

Table 1. Summary of estimates of water vapour feedback (and one of lapse rate) from interannual variability. LW= long wave feedback, SW= short wave feedback, Net=LW+SW. Range of values shown is ± 2σ.

The range shown in Table 1 is unsurprising, given the diversity of approaches adopted across the studies. Some consider different phase strong ENSO events alone (Dessler and Wong, 2009) while others perform regressions of radiative changes with monthly (Dessler, 2013; Gordon et al., 2013; Liu at al., 2018) or annual (Colman and Hanson, 2012) temperature across multiple years irrespective of ENSO activity. Regression methodology, too, can make significant differences, and in particular explain much of the very large variation in feedback strength found between related studies by Dessler and Wong (2009) and Dessler (2013). Differences can also arise because of choice of radiative kernels used (Liu et al., 2018) along with one study applying the PRP technique (Colman and Hanson, 2012, see Appendix 1). Estimates using reanalyses must also contend with shortcomings in their representation of radiatively important field such as moisture over dry subtropical ocean regions in some reanalysis datasets (Allan et al., 2004), and substantial differences between the reanalyses themselves such as numerical weather prediction model base and assimilated datasets (Koumoutsaris, 2013).

All but one study (Dessler, 2013) evaluated only traditional water vapour feedback (Eqn 4). Using the alternate fixed relative humidity formulation (Held and Shell, 2012; see Eqns 6-8), Dessler (2013) found, as expected, that the disagreement between results from two reanalysis decreased – giving common values of λ'_P term of -1.92 $Wm^{-2}K^{-1}$, and the λ'_H term of ~-0.06 $Wm^{-2}K^{-1}$. However not all disagreement disappeared, with the λ'_Γ term varying significantly from 0.09 to 0.26 $Wm^{-2}K^{-1}$ (Dessler, 2013).

A key remaining difference between the studies is their choice of time periods for feedback evaluation. The effect of this can be very great and poses an additional challenge for determining interannual water vapour feedback from observations. Liu et al. (2018) show that sampling 19 different 12-year segments (corresponding to the length of the available AIRS-MLS dataset) from 30-year periods from CMIP5 models give widely varying estimates of feedback strength (see Fig. 24). Further, the mean of the 19 samples could differ strongly from results from the whole 30-year period (Fig. 24). An investigation of five different 20-year samples over the full 20th century from CMIP3 models (Dalton and Shell, 2013) backs this up. It found substantial variation in the analysed interannual variability feedback from over the 5 samples, although despite this it showed modest cross model correlation between variability derived water vapour feedback and secular water vapour feedback over the full century (Dalton and Shell, 2013). This high sensitivity to time sampling in calculating interannual water vapour feedback needs to be borne in mind, for example, when considering values such as from Gordon et al. (2013) which are based on only 88 months of observations (which was the limit of the AIRS data available).

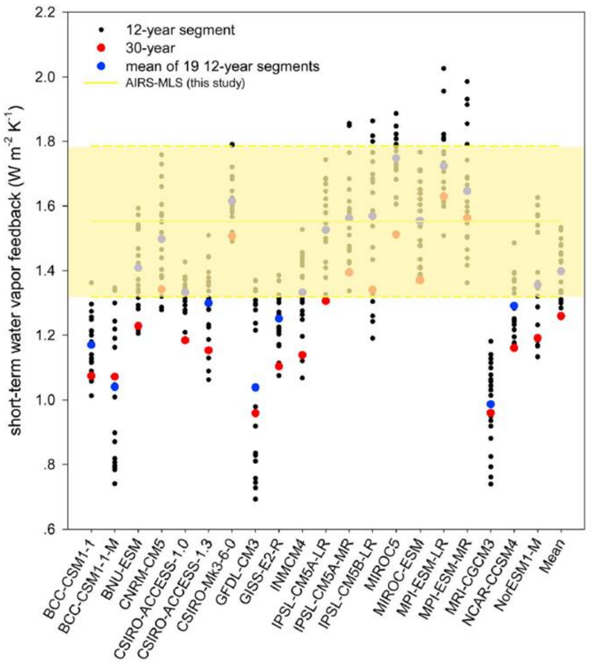


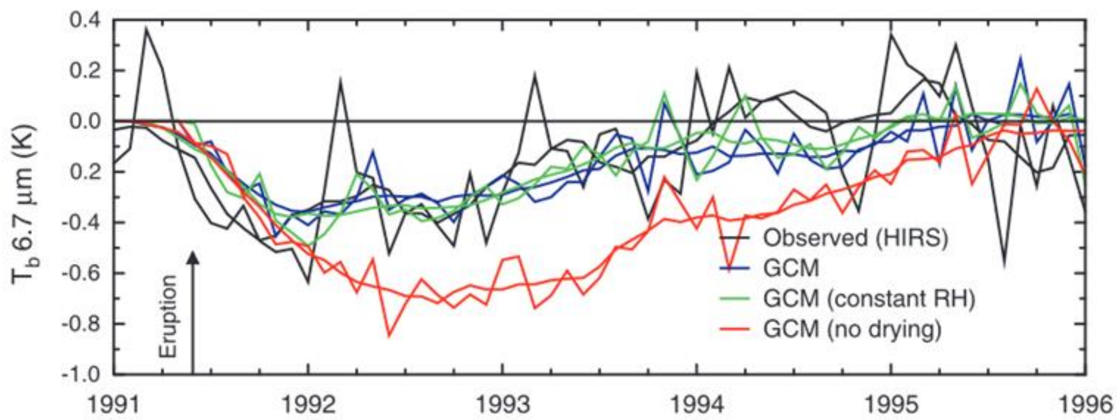
Figure 24. (Colour online) Water vapour feedback calculated from interannual variability from CMIP5 GCMs forced by observed SSTs over the period 1979-2008. Red (larger) dots show feedbacks calculated over the entire 30 year period, black dots from 19 different 12 year segments within this period, and blue dots the mean of all 12 year segments. Shading shows AIRS-MLS sounding observations from 2004-2016. It is clear that an enormous spread of results from estimation of water vapour feedback can be expected if taken from relatively short (~decadal) time periods of models or observations. Source: Liu et al. (2018). Reproduced with permission from the American Geophysical Union.

In summary then, seasonal, interannual and longer timescales provide important information and tests on water vapour feedback, although do not provide direct analogues for climate change feedback. There remains considerable observational uncertainty in the value of interannual water vapour feedback, and there are major challenges in refining it. However, if the range of estimates could be better understood and narrowed, potential constraints may be possible using correlations between interannual (and longer term) feedbacks and feedbacks under secular climate change (Gordon et al., 2013; Colman and Hanson, 2016; Takahashi et al., 2016; Liu et al., 2018; Dalton and Shell, 2013).

D. Volcanoes and water vapour feedback

Volcanic eruptions provide another potential analogue for long-term climate change. Large explosive volcanoes can emit vast quantities of aerosols into the stratosphere, blocking sunlight for extended periods causing a multi-year transient cooling of order a few tenths of a °C (Robock and Mao, 1995; Kirtman et al., 2013). The most recent large, climatologically significant eruption, that of Mount Pinatubo in 1991, provides a "natural experiment" for testing water vapour feedback and its role in climate response.

Observed 6.7μm channel radiances (sensitive to relative humidity averaged over roughly 200 to 500 hPa), show modest reductions in the years immediately following the eruption (black line in Fig. 25a, taken from Soden et al., 2002). GCM simulated emission temperatures show good agreement with observations (blue and green lines) either calculated directly from the model, or else assuming constant relative humidity. However, they show roughly doubled observed 6.7μm emission temperature reduction if no relative humidity induced drying occurred (red line). Together this provides direct evidence that roughly unchanged upper tropospheric relative humidities occur under the cooling found in response to Pinatubo (Soden et al., 2002).



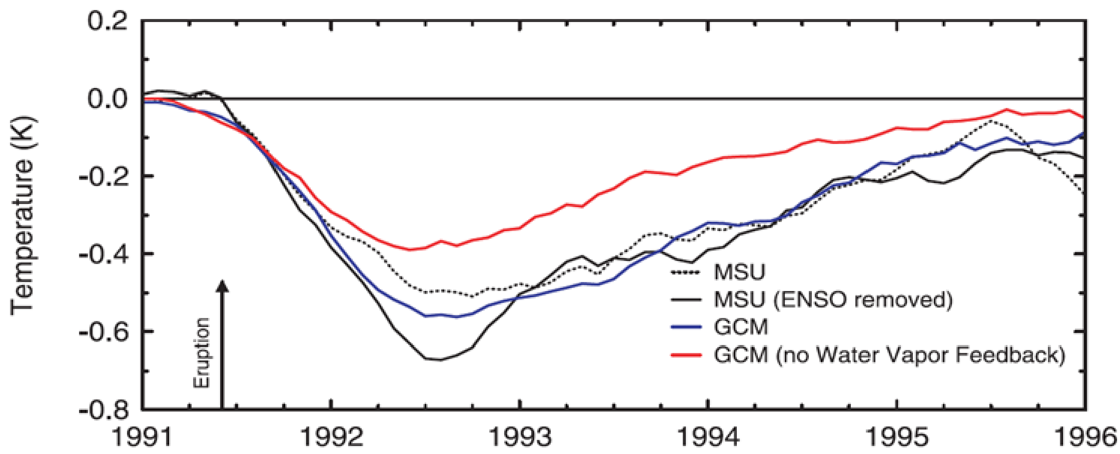


Figure 25. Top: Changes in global mean 6.7μm brightness temperatures from High Resolution Infrared Radiation Sounder (HIRS) observations (black line) and as simulated by the GFDL GCM

(blue line) following the Mount Pinatubo eruption. Anomalies are expressed relative to pre-eruption values (January-May 1991). The green line represents the GCM with unchanged *relative* humidity, and the red line with unchanging *specific* humidity (corresponding to no upper tropospheric drying). Thick lines are seven month running means. This shows that unchanging relative humidity, not specific humidity, enables the model to match observations.

Bottom: Observed global cooling in the lower troposphere following the Pinatubo eruption as measured by the Microwave Sounding Unit (MSU) (black lines) and model predicted temperature (blue line). The red line shows the temperature trace of the GCM with suppressed water vapour feedback. This shows that strong positive water vapour feedback was necessary for the GCM to reproduce observed cooling. From: Soden et al. (2002). Reprinted with permission from AAAS.

To investigate the impact of water vapour feedback on the *global cooling* that followed, Soden et al. (2002) used a modified GCM which suppressed the terrestrial radiative response to water vapour changes. They showed that whereas an unmodified version of the GCM was able to represent the magnitude of post Pinatubo global cooling (peaking at around -0.5° C), the GCM without water vapour feedback produced much weaker cooling (Fig. 25b).

An extension by Forster and Collins (2004) considered the vertical and meridional distribution of water vapour changes in response to Pinatubo using satellite-derived vertical water vapour observations. Calculating the radiative effect, they estimated a value of the water vapour feedback of 1.6 $Wm^{-2}K^{-1}$, with a 5-95% range from 0.9 to 2.5 $Wm^{-2}K^{-1}$. Parallel calculations from a large model ensemble forced by stratospheric aerosol observations produced a comparable average of 2.0 $Wm^{-2}K^{-1}$ (range 0.4 to 3.6 $Wm^{-2}K^{-1}$). Caution must be exercised in interpreting the results. No vertical profile from the GCM ensemble closely matched the observations, and natural variability unrelated to the volcanic forcing caused considerable spread in both observational and model results (Forster and Collins, 2004).

Together these studies provide compelling evidence for strong positive water vapour feedback following climate forcing (Del Genio, 2002). Of course, some differences in water vapour feedback strength may be possible from volcanic aerosol forcing as distinct from GHG forcing. However, GCM experiments considering both volcanic aerosol and CO_2 forcings find only small differences in the net clear sky response (Yokohata et al., 2005) or in the water vapour feedback itself (Yoshimori and Broccoli, 2008).

E. Paleo evidence

Paleo climates provide another line of evidence on the magnitude of water vapour feedback. Paleo reconstruction and modelling evidence indicates that a strong positive water vapour feedback is needed to explain both colder (Berger et al., 1993; Crucifix, 2006) and warmer (Lariviere et al., 2012) paleo climates. For example, 2D modelling sensitivity studies by Berger et al. (1993) found that water vapour feedback was responsible for around 40% of the cooling during the LGM. Strong positive water vapour feedback has also been found to help explain impacts of LGM continental ice sheet thicknesses on high latitude temperatures (Liakka and Lofverstrom, 2018).

A study using six CMIP5-PMIP3 (Paleo Model Intercomparison Project phase 3) models found that water vapour feedback was responsible for around 29% of the global cooling during the Little Ice Age, 1600-1850 CE (Atwood et al., 2016).

A consideration of reconstructed 800,000-year temperatures from ice-core data (across multiple glacial/interglacial cycles) shows self-consistency for climate sensitivity of 3K, consistent with strong positive net water vapour+lapse rate feedback (Hansen et al., 2008; Lacis et al., 2013).

Together these studies provide important, albeit indirect, supporting evidence for a strong, positive water vapour (and water vapour+lapse rate) feedback, which has acted to amplify past climate change.

F. Observed lapse rate changes and variability

The section has so far concentrated largely on water vapour changes. Since lapse rate feedback is dominated by tropical changes in saturated adiabatic lapse rate, we would expect to observe warming in the upper troposphere to have exceeded surface warming over recent decades. However, there has been controversy over the past 30 years on observed global or tropical mean lapse rate changes, starting with suggestions in the early 1990s that models have overdone upper tropospheric temperatures increases compared with observed changes (Spencer and Christy, 1990). The implication was that models may be missing or misrepresenting processes driving lapse rate responses under projected climate change. Long debate has taken place about the significance and cause of differences, including uncertainties in the observations (Flato et al., 2013).

Behind much of this uncertainty has been that both satellite and radiosonde observations are characterised by time varying biases and discontinuities (Po-Chedley et al., 2015), with the radiosonde network also featuring regional inhomogeneities and large data sparse regions. For example, producing long-term tropical timeseries from advanced microwave sounding unit (AMSU) measurements has been challenging, with different estimates depending on factors such as treatment of different satellites and diurnal cycle corrections (Po-Chedley et al., 2015). Removing temperature biases has also proven sensitive to methodology (Thorne et al., 2011).

There is evidence that tropical upper tropospheric warming in GCMs overall have exceeded observations of the last several decades (McKitrick and Christy, 2017; Santer et al., 2017), although some of the CMIP5 models agree with the observations within error estimates (Flato et al., 2013). Decadal timescale variability may explain some of the disagreement, but deficiencies in the forcing applied to models, such as from volcanic eruptions (Santer et al., 2014) or from other atmospheric aerosol changes (Santer et al., 2017) have been found to also contribute. Despite possible observational/model disagreements, however, observed trends in mid-upper tropospheric temperatures (Santer et al., 2013), and trends in the seasonality of atmospheric temperatures (Santer et al., 2018) are incompatible with natural variability alone, indicating a human influence.

Patterns of SSTs changes also influence the strength of lapse rate feedback diagnosed from observations. Warming, as has occurred in the observed trend (enhanced West Pacific compared with eastern Pacific warming, Hartmann et al., 2013) results in a stronger negative lapse rate feedback in models than for more uniform warming such as under equilibrium (i.e. long-term) climate change (Andrews and Webb, 2018). This "unexpected" pattern of warming is important as it also has impacts on net climate sensitivity as it affects Pacific-wide cloud changes, particularly for low cloud in the east (Andrews and Webb, 2018).

Studies using models forced by observed SSTs (rather than fully coupled models) provide a promising "cleaner" comparison with observed upper tropospheric warming (Mitchell et al., 2013). Model results can vary between different SST datasets, however, so possible observational SST errors add further uncertainty (Flannaghan et al., 2014). Furthermore, results can differ substantially between models for a given SST dataset due to differing precipitation pattern responses (Fueglistaler et al., 2015).

A recent study of Coupled Model Intercomparison Project phase 6 (CMIP6, Eyring et al., 2016) models showed that, as in the two previous model generations (Fu et al., 2011; Po-Chedley and Fu, 2012), there remains an overestimate of upper tropospheric warming compared to observations—in this case radiosondes and ECMWF reanalyses (Mitchell et al., 2020, Fig. 26). However, much of the overestimate can be linked to biases in surface temperature increases, rather than lapse rate change deficiencies, as tropospheric temperature agreement is much closer when models are forced by observed SST changes (see Fig. 26).

Confidence in models is also reinforced by *variability* studies. On monthly to interannual timescales, a range of observations including radiosondes and MSU satellite observations show an amplification of warming with altitude, in a manner that agrees with theory and climate model simulations (Santer et al., 2005).

Further details of the debate on upper tropospheric warming in models versus observations are beyond the scope of this review paper. An extensive review, albeit not recently updated, is provided by

Thorne et al. (2011) and readers are referred to the Fourth and Fifth IPCC Assessment Reports (Hegerl et al., 2007; Hartmann et al., 2013) for further discussion.

Regardless, the offsetting nature of temperature and water vapour responses in the tropical upper troposphere means that *combined* tropical water vapour plus lapse rate feedbacks are insensitive to such differences (Boucher et al., 2013; Ingram, 2013a, b; Po-Chedley et al., 2018). Therefore, the uncertainties of tropical lapse rate changes, although important for understanding the individual feedbacks and the representation of processes in models, do not significantly decrease confidence in the strength of combined water vapour and lapse rate feedbacks in GCMs.

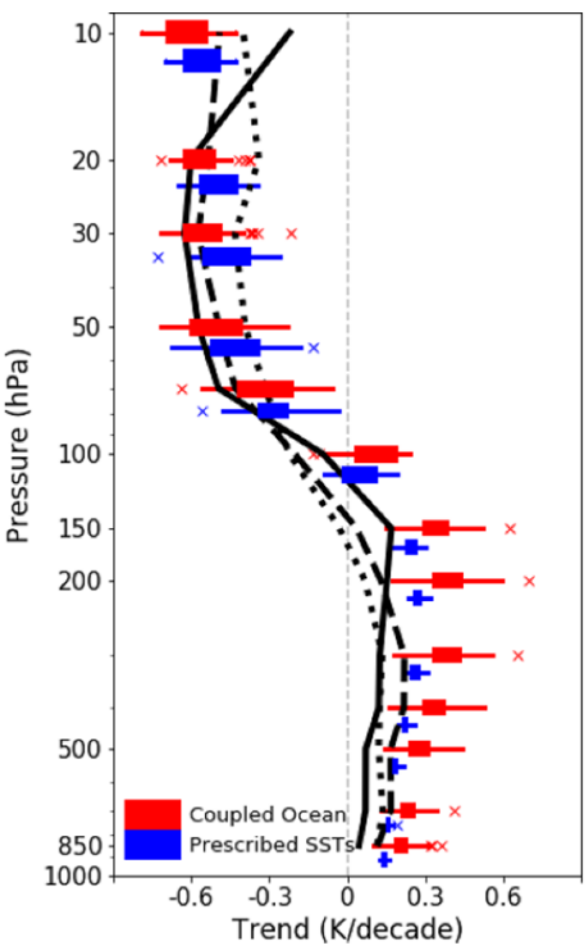


Figure 26. (Colour online) Comparisons of CMIP6 models vertical temperature trends (20° N/S) with observations from two radiosonde datasets (RICH1.7 and RAOBCORE1.7) and the ERA5/5.1 reanalysis (black lines), for the period 1979-2014. Box-whisker plots show the 25-75% inter-model range, bars and crosses represent the 1.5 quartile range then outliers beyond this. Red (upper) boxes represent 48 fully coupled ocean/atmosphere CMIP6 GCMs forced by anthropogenic greenhouse gases and aerosols as well as estimated "natural" forcing, and blue (lower) boxes a subset of 28 atmospheric only GCMs forced by observed SST changes. Blue (lower) lines are displaced vertically for plotting purposes.

In the troposphere, models show much less spread, and better agreement with observations when forced by SSTs that match observed, rather than those simulated under a "freewheeling" experiment, indicating that much of the apparent disagreement between model and observed lapse rate changes may be due to different trends in model surface temperatures. Source: Mitchell et al. (2020).

VI. Model representation of feedbacks and feedback processes

The previous section described the extensive observational evidence supporting water vapour and lapse rate feedbacks. Models, in combination with observations, are fundamental to our understanding of feedback processes and for providing quantitative estimates of their strength. Crucially too, models represent our primary tool for projection of future climate change. Extensive research, presented in this section, has therefore focussed on the evaluation and assessment of models and model processes underpinning water vapour and lapse rate feedbacks.

A. Quantification of feedbacks in models

Limitations in the understanding and evaluation of water vapour and lapse rate feedbacks in GCMs through the 1990s and early 2000s in part related to challenges in their quantification in models and observations. A critical advance in this area in the last 40 years and particularly over the last 20 has been the development of methods of calculating and comparing model feedbacks, and determining feedbacks from observations. These methodologies are described in Appendix 1.

B. Model representation of water vapour distribution, variability, and trends, and their radiative impact

For the mean climate, models represent with skill large-scale features of the observed relative humidity field, and associated OLR (Bates and Jackson, 1997; Gaffen et al., 1997; Randall et al., 2007; Flato et al., 2013). For example, comparison with observations from the Atmosphere Infrared Sounder (AIRS) showed that CMIP5 models overall represented distributions of tropospheric specific humidity and temperature well (Tian et al. 2013). There were, however, some notable biases, including a cold bias of around 2° C in the extratropical upper troposphere, and a moist bias in the tropical upper troposphere (Tian et al. 2013). Comparison of the CMIP5 model specific humidity with NASA A-train moisture retrievals show agreement to within 10% in the low to mid troposphere (Jiang et al., 2012). In the upper troposphere, however, a larger range is found in models, from around 1% to twice the observed value (Jiang et al., 2012). This represented limited advance from the earlier generation of CMIP3 models, which on average had a bias of over 100% in free tropospheric specific humidity (John and Soden, 2007; Jiang et al., 2012). Further modest improvement has been found in CMIP6 compared with CMIP5 (Jiang et al.). Although some of these biases remain substantial, *fractional* change in water vapour rather than *absolute* change is critical for the feedback. Therefore, such "present climate" biases should not be crucial to net feedback strength (Held and Soden, 2000; John and Soden, 2007) and indeed, biases in the current climate are uncorrelated with the magnitude of water vapour feedback (John and Soden, 2007).

The observed tropical "bimodality" in the humidity distribution is represented with widely varying skill in GCMs (Zhang et al., 2003; Pierrehumbert et al., 2007). Bimodality is indicative of sharp moisture gradients, and parcel mixing timescales being longer than moisture residence times in the tropical atmosphere (Zhang et al., 2003). However, the importance of this feature on feedback processes and the representation in models is unclear (Randall et al., 2007), and no evidence has established that this issue adversely affects model representation of water vapour feedback.

Although, as discussed above, feedback strengths under interannual or decadal variability are not direct analogues for secular climate change feedback, skilful representation of observed variability can nevertheless bolster confidence that models represent key processes controlling upper tropospheric humidity on these timescales, and under these temperature forcings (Randall et al, 2007).

Studies show that models can reproduce observed interannual variations in *lower tropospheric* moisture, itself consistent with approximately invariant relative humidity (Soden and Schroeder, 2000; Allan et al., 2003, Trenberth et al., 2005) – see Section V-A. This is an important test for model representation of moisture variability generally, but is unsurprising, given the tight coupling between surface and lower troposphere, and the wide-spread availability of surface water (Bony et al., 2006).

For the *upper troposphere*, models show skill in representation of OLR and observed water vapour variations from seasonal changes (Inamdar and Ramanathan, 1998; Tsushima et al., 2005). Many studies have also found overall model skill in representing interannual moisture variations and associated radiation changes (Soden, 1997; Kiehl et al. 1998; Soden, 2000; Dessler and Sherwood, 2000; Gettelman and Fu, 2008; Dessler and Wong, 2009). For example, models reproduce interannual water vapour feedbacks derived from reanalysis temperature and moisture data (Slingo et al., 2000; Dessler and Wong, 2009; Colman and Hanson, 2012; Dessler, 2013), and show a modest decrease in relative humidity with temperature within the error bars of observations at 215 hPa (Minschwaner et al., 2006). A recent study found tropical mean 200 hPa specific humidity variations with temperature measured by three methodologies—GPS refractive indices, AIRS satellite retrievals and the MERRA reanalysis—to lie well within the range simulated by CMIP5 models, albeit a little below the multi-model mean (Vergados et al., 2016; Fig. 23).

Figure 22 shows TOA radiation perturbations due to water vapour changes over ENSO events for 12 CMIP3 models and two reanalyses. It confirms that models represent interannual fluctuations in moisture and radiation response similarly to estimates from observations, in both meridional and vertical dimensions. CMIP3 models have also been found to straddle two reanalysis estimates for both interannual and seasonal water vapour feedback (Colman and Hanson, 2016) and 20 CMIP5 models with values calculated using Atmospheric Infrared Sounder and Microwave Limb Sounder satellite observations from 2004-2016 (Liu et al., 2018). Feedbacks from fully coupled models, however, are on average slightly weaker than those from models forced by observed SSTs (Liu et al., 2018).

Critically, models also show *trends* of upper tropospheric humidity consistent with satellite observations over the period 1982 to 2004 (Soden et al., 2005). Results using satellite "emulators"[16] within models of upper tropospheric humidity dependent radiances such as of HIRS 14 μm wavelength, find model skill in representation of interannual and decadal variability and long-term trends (Allan et al., 2003). Similarly, a more recent study found CMIP5 models overall reproduced satellite derived tropical upper tropospheric humidity trends over 27 years ending 2005 (Chung et al., 2014).

In summary, then, models show significant skill in reproducing observed trends and variability of relative humidity in both the upper and lower troposphere, and consequently of water vapour feedback under secular change and interannual variability. Models overall have skill in representing large scale mean distributions of humidity, but with biases in some regions. Because of the logarithmic dependence of radiation changes on specific humidity however, these biases do not affect model estimates of water vapour feedback. Together these findings strongly reinforce confidence in model representation of water vapour feedback.

C. Conclusions on feedback impacts on global variability.

Apart from observations of variability providing exacting tests for models, studies presented in the previous section and Section V-B provide overwhelming evidence that water vapour feedback (and combined water vapour+lapse rate feedback) *amplify global temperature variability* across a wide range of timescales.

Observations and models confirm that water vapour feedback reinforces the annual cycle (Hu, 1996; Tsushima et al., 2005; Wu et al., 2008), with large positive values in individual summer hemispheres (Colman and Hanson, 2013; Fig. 21), although with relatively weak annual mean values because of strong seasonal hemispheric offsetting. On more limited evidence, models also suggest that lapse rate feedback, too, amplifies both global (Colman and Hanson, 2013; Fig. 21) and mid latitude (Hu, 1996) seasonal cycles.

Models and observations are also unanimous on the amplification of interannual global temperature variability (see discussion in Sections V-C and VI-B). Apart from the overwhelming diagnostic evidence presented in these sections, direct evidence comes from suppressed water vapour feedback experiments in models, which find that removal of water vapour feedback decreases unforced interannual temperature fluctuations (Hall and Manabe, 1999, 2000a). Lapse rate feedback impact on interannual variability has been evaluated more rarely, with findings that it remains a negative feedback thereby suppressing interannual variability (Colman and Hanson, 2013, Koumoutsaris, 2013; Fig. 21).

More limited evidence on *decadal* variability indicates that water vapour also increases temperature variations on these timescales (Allan et al., 2003; Colman and Hanson, 2013; Colman and Power, 2018), but that lapse rate feedback dampens them (Colman and Hanson, 2013, 2016). Limited available evidence also suggests decadal feedback is modestly weaker in magnitude on average than under long-term climate change (Fig. 21).

[16] An emulator is model code that simulates the radiances as they would be directly "seen" by a satellite.

VII. Evaluation and assessment of water vapour and lapse rate feedbacks.

A. Intergovernmental Panel on Climate Change (IPCC) assessment of water vapour and lapse rate feedbacks

The IPCC has made prominent, and oft quoted, assessments of the magnitude of, and overall confidence in, water vapour and combined water vapour+lapse rate feedbacks. The importance of the IPCC assessments is that they strive to represent a *climate community-wide* evaluation of relevant evidence. The IPCC process facilitates this through a broadly representative lead and contributing author list, and three stages of expert, community and government reviewing (Interacadamy Council, 2010). Table 2 lists the "headline" assessments of the First through Sixth Assessment Reports. The IPCC has only occasionally explicitly estimated the value of water vapour/lapse rate feedbacks, instead providing qualitative or semi-quantitative statements about their strength, the ability of models to faithfully represent key processes, as well as overall assessed confidence.

Report	Assessment
FAR, Cubasch and Cess (1990)	The best understood feedback mechanism is water vapor feedback, and this is intuitively easy to comprehend
Supplementary report, Gates et al. (1992)	There is no compelling evidence that water vapor feedback is anything other than positive—although there may be difficulties with upper tropospheric water vapor
SAR: Dickinson et al. (1995).	Feedback from the redistribution of water vapor remains a substantial source of uncertainty in climate models—Much of the current debate has been addressing feedback from the tropical upper troposphere, where the feedback appears likely to be positive. However, this is not yet convincingly established: much further evaluation of climate models with regard to observed processes is needed. Changes in lapse rate act as an additional feedback that can also be substantial and that generally oppose the water vapour feedback.
TAR: Stocker et al. (2001)	Models are capable of simulating the moist and very dry regions observed in the tropics and subtropics and how they evolve with the seasons and from year-to-year…. While reassuring this does not provide a definitive check of the feedbacks, although the balance of evidence favours a positive clear-sky water vapour feedback of the magnitude comparable to that found in simulations.
AR4: Randall et al. (2007).	New evidence from both observations and models have reinforced the conventional view of a roughly unchanged relative humidity response to warming…. Taken together, the evidence strongly favours a combined water vapour-lapse rate feedback of around the strength found in GCMs.
AR5: Boucher et al. (2013)	The net feedback from water vapour and lapse rate changes combined, as traditionally defined, is extremely likely (more than 95% confidence) positive… Values in this range [0.9-1.3 $Wm^{-2}K^{-1}$] are supported by a steadily growing body of observational evidence, model tests and physical reasoning.
AR6: Forster et al. (2021)	The combined water vapour plus lapse rate feedback is positive. The main physical processes that drive this feedback are well understood and supported by multiple lines of evidence including models, theory and observations. The combined water vapour plus lapse rate feedback parameter is assessed to be 1.30 $Wm^{-2}K^{-1}$, with a very likely range of 1.1 to 1.5 $Wm^{-2}K^{-1}$ and a likely range of 1.2 to 1.4 $Wm^{-2}K^{-1}$ with high confidence.

Table 2. IPCC assessments of water vapour and lapse rate feedbacks, for the First (FAR), Second (SAR), Third (TAR), Fourth (AR4), Fifth (AR5) and Sixth (AR6) Assessment reports and for the Supplementary Report.

A close examination of Table 2 is instructive on the progress of confidence in water vapour and lapse rate feedbacks over the last 30 years.

In the First Assessment report (1990) it was deemed self-evident that water vapour feedback was strong and positive. However, the scientific challenges to the mainstream view in the early 1990s (see Section IV-D) pointing out the critical nature of the (then, poorly understood) tropical upper tropospheric moisture changes, and the central role of convective detrainment and associated uncertainties (Lindzen, 1990; Lindzen, 1994) caused a re-examination of confidence in the nature and strength of the feedback. This resulted in the much more ambivalent statements in the Supplementary Report (1990) and the Second Assessment Report (1995), featuring an emphasis on poorly understood processes governing upper tropospheric humidity changes.

An accumulation of research on theory, processes, modelling and observational studies in subsequent years led to a steady increase in confidence in the sign and strength of the combined feedbacks through the Third, Fourth, Fifth and Sixth Assessment Reports. It pays to now reflect upon a timely challenge to those arguing for weak or negative water vapour feedback, issued by the Third Assessment Report (Stocker et al., 2001). That challenge was to develop a GCM that reproduces observed climate, and yet has a substantially weaker water vapour feedback than contemporary GCMs. No such model has ever been produced.

B. Summary "best" estimate of feedback strengths and uncertainty ranges.

Many studies have provided estimates of water vapour, lapse rate and combined water vapour plus lapse rate feedbacks, either from models or derived from observations. A list of estimates, including the methodology and references are provided in Table A2-1 in Appendix 2. A summary of feedback estimates is provided in Fig. 27.

An early estimate was made from the IPCC First Assessment Report (Cubasch and Cess, 1990) based largely on two published studies by Cess et al. (1989), and Raval and Ramanathan (1989) that considered observations of temperature sensitivity of OLR, as well as modelling studies from 14 GCMs under simplified forcing. That estimate (1.2 $Wm^{-2}K^{-1}$) now sits well below the range of subsequent estimates. Other IPCC Assessment Reports have generally not provided evaluations of the magnitude of water vapour or lapse rate feedback separate to those in the literature, apart from the AR5 which provided an estimate of the combined feedback of 1.1 (0.9 to 1.3, 90% range) $Wm^{-2}K^{-1}$.

Purely modelling based estimates of feedback strength have a long history. A study by Colman (2001) assembled already published model results from RCMs and GCM studies, and derived a multi-model combined feedback estimate of 1.37±0.4 $Wm^{-2}K^{-1}$. This is a higher number than other estimates – e.g. from the AR5, although it included a diverse range of models including RCMs. Other multi-model estimates have typically been calculated using radiative kernels applied to a range of CMIP experiments (see Table A2-1). As a result, we have estimates from CMIP2, 3, 5 and 6 ensembles (Soden and Held, 2006; Koumoutsaris, 2013; Vial et al., 2013; Caldwell et al., 2016; Colman and Hanson, 2017; Zelinka et al., 2020). It is noteworthy that there have been high levels of overall consistency down the years between different model generations, despite two decades of model development, and a dramatic increase in horizontal and vertical resolution, albeit with the suggestion of slightly stronger water vapour feedback in the last two generations of models (Table A2-1 and Fig. 27). It is also notable that overall consistency between generations holds despite a significant increase in ECS from CMIP5 to CMIP6, including some models with sensitivities of over 6K (Zelinka et al., 2020).

A caveat on such comparisons is that different climate change experiments produce slightly different feedback strengths, as can be seen by comparing water vapour and lapse rate feedback estimates for CMIP5 from Historical, abrupt 4xCO_2 and RCP 8.5 projection experiments (Colman and Hanson, 2016). This is unsurprising given the sensitivity of water vapour and lapse rate feedbacks to SST warming patterns, as discussed in section IV-G. Furthermore, different kernels can produce somewhat different feedback strengths (Vial et al., 2013).

Another approach to estimating climate change feedbacks has been to use observations to estimate interannual feedback, then employ model derived correlations between interannual and climate change feedbacks. For example, this method was used by Gordon et al. (2013) and Liu et al. (2018) to estimate climate change feedback using AIRS-MLS based interannual feedback measurements. The appeal of this approach is that it has a basis in observations rather than models alone. The disadvantages are that the relatively short time periods available for the observations inevitably result in significant sampling uncertainties in the strength of the feedback (Section V-C), and the technique also relies on the validity and accuracy of the correlation between interannual and climate change feedbacks in GCMs. Finally, another observational approach is that of Forster and Collins (2004), who used the cooling following the Pinatubo eruption to estimate a water vapour feedback of ~1.6 $Wm^{-2}K^{-1}$.

A recent review of variations in ECS across models by Sherwood et al. (2020) compared water vapour, lapse rate, surface albedo and cloud feedbacks across models and observations. This included estimates from two separate model ensembles CMIP 5 (Taylor et al., 2012) and CMIP 6 (Eyring et al., 2016), as well as estimates from interannual variability using ERA Reanalyses (Dessler, 2013).

Sherwood et al. (2020) demonstrate exhaustively that uncertainty in cloud feedbacks remains the biggest source of uncertainty in model ECS. The final estimate of water vapour plus lapse rate feedback strength made by Sherwood et al. (2020) is 1.15 ± 0.15 $Wm^{-2}K^{-1}$ (the range representing one standard deviation), which compares closely with the estimate from the IPCC AR5 of 1.1 (0.9 to 1.3, 90% range) $Wm^{-2}K^{-1}$. On top of this we need to consider possible stratospheric water vapour feedback. Estimates of this are much fewer and contain considerable uncertainty (Section IV-F), with a recent calculated CMIP5 range of 0.10 to 0.26 $Wm^{-2}K^{-1}$ (Banerjee et al., 2019), although compensating temperature feedbacks may largely offset much of this (Section IV-F).

Our best estimate of overall strength of combined water vapour+lapse rate feedbacks is 1.25 ± 0.15 $Wm^{-2}K^{-1}$ (the range being one standard deviation), based on expert judgement from the range of results in the literature. This is a value slightly larger than that of Sherwood et al. (2020) and Boucher et al. (2013), taking consideration of the higher feedback strength from the last two generations of CMIP models (Fig. 27) and the evidence for at least a modest positive stratospheric contribution. It is very close to (marginally below) the estimate of Forster et al. (2021).

The RH-based feedback approach (Section III-B) provides a second and related perspective on this estimate. Adding the component of the "Planck" term in Eqn 6 that corresponds with the radiative effect of the humidity increase from uniform warming to the modified lapse rate and relative humidity feedbacks equals the traditionally defined water vapour+lapse rate feedback.

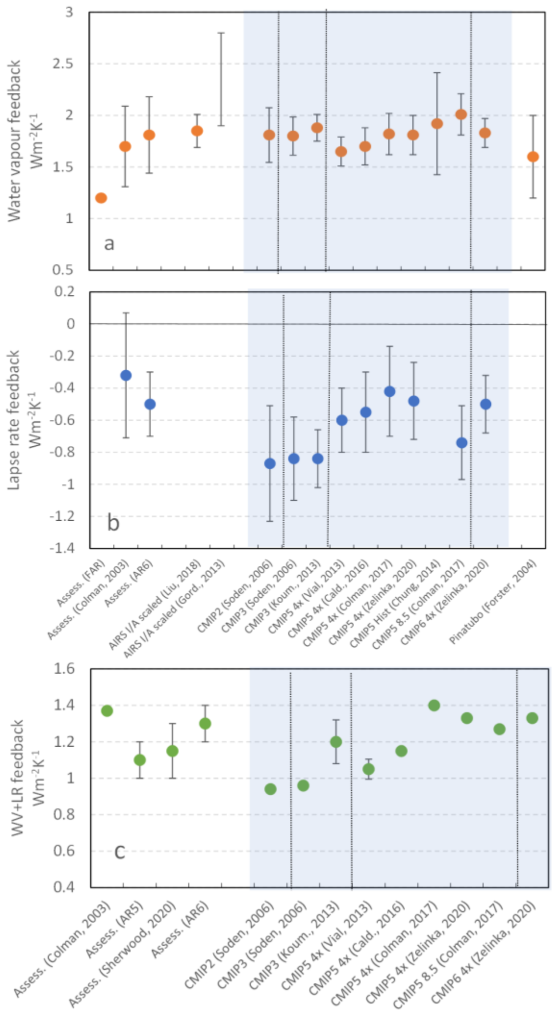


Figure 27. (Colour online) Values of (a) water vapour, (b) lapse rate, (c) combined water vapour plus lapse rate feedbacks taken from the studies listed in Table A2-1. Error bars show ±1σ range of the estimates (where available). "Assess."

refers to an evaluation carried out from published literature. Experiments referred to are "4x", 4xCO_2; "Hist", CMIP Historical simulations; "8.5", CMIP RCP8.5 experiments. Shaded areas denote CMIP-based analyses, with the vertical lines differentiating CMIP2, 3, 5 and 6. "AIRS I/A scaled" are determinations from Atmospheric Infrared Sounder observations of interannual feedback, scaled by the ratio between climate change and interannual feedbacks derived from model ensembles. AR6 water vapour values are the average of the reported model and observational estimates.

VIII. Conclusions

A. On the strength and consistency of evidence for water vapour and lapse rate feedbacks

The last three decades have seen enormous progress in understanding of water vapour and lapse rate feedbacks, in their observation, and the understanding and representation of processes in climate models. This research has transformed our understanding of these feedbacks, and established beyond doubt that the water vapour feedback (as traditionally defined, see Section III-A) operates as a strong positive feedback in the climate system, on its own roughly doubling the response to GHGs forcing compared with a hypothetical climate system without feedbacks, i.e. with Planck cooling alone. This research has also established that lapse rate feedback is a moderate negative climate feedback and essentially beyond doubt that the *combined* water vapour plus lapse rate feedback is a moderate to strong positive feedback in the climate system.

For confidence in water vapour feedback, understanding of processes controlling humidity distributions and their change under global warming is critical. Overall, these physical processes are now well understood. Theoretical understanding points to overall unchanged relative humidity with warming. Further, the "partly-Simpsonian" explanation of the spectral dependence of TOA radiation on surface temperature has provided a firm theoretical basis for understanding the water vapour feedback.

Large-scale humidity structures in the mid to upper tropical troposphere can be traced to detrainment and mixing from convection advected by winds, along with moisture mixed in from mid-latitude intrusions. To first order, broadscale relative humidity is unchanged under global warming, although in closer detail there are widespread modest projected relative humidity changes (~1-2%/K of warming), including decreases in the tropical upper troposphere and at mid-latitudes, with increases in the tropical lower troposphere. These modest decreases in the mid to upper troposphere are found both in models and observations and weaken water vapour feedback by roughly 5% compared to fixed relative humidity.

The broadscale humidity distribution, and its change with temperature is not sensitive to uncertainties stemming from model convection or cloud microphysics. This is evidenced by similar water vapour feedback being found in multiple generations of models with widely varying resolution and physical parameterisations: no GCM described in the peer-reviewed literature has ever been constructed with small or negative water vapour feedback. Secondly it has been convincingly demonstrated that the broad scale observed and modelled humidity distributions can be well represented by "Advection-Condensation" models which eschew microphysics altogether but instead proscribe only conservation of last saturation humidity sourced from rising convective regions, then advection by broadscale winds. The large-scale organisation of the atmosphere into concentrated rising regions rather than widespread small-scale convective cells appears to be central to this insensitivity to convective/cloud microphysics. On top of this there is no evidence of significant moistening resulting from evaporation of advective precipitated water – e.g. from clouds, further reinforcing that the details of cloud microphysics are not important for overall humidity distribution.

This coherent view, combining theory, observations and simple and complex modelling studies establishes the confidence in water vapour providing a strong positive feedback, and it being of about the strength found in GCMs. Other lines of evidence in support come from the observed and modelled response to volcanic eruptions, and from paleo reconstructions which provide evidence that a strong positive water vapour feedback is required to explain global and regional temperature changes in past warmer and colder climates. Although not an exact analogue for climate change, the close-to-unchanged relative humidity observed and modelled on large scales under natural variability further reinforces the picture.

New observational datasets, and modelling studies suggest there may be significant contributions to water vapour feedback from the stratosphere, through temperature dependent penetration of moisture from the mid latitude in tropical upper troposphere into the lower stratosphere. However, debate continues regarding dominant processes, and on whether the overall impact on the climate is one of warming once compensating temperature related feedbacks are taken into consideration in the upper troposphere and lower stratosphere.

Another major advance over the last 20-30 years is the appreciation that water vapour feedback also amplifies variability on seasonal, interannual and decadal timescales. Estimates of the strength of interannual feedback vary substantially across models, and from differing observations. The reason for the spread in the latter is likely because of different methodologies, use of different data sets, such as different reanalyses or satellite products, different selected time periods and the shortness of sampling. Differences in surface temperature patterns associated with variability and long-term climate change means that measurements of feedback under interannual variability, for example, are not direct analogues of water vapour feedback under climate change. Nevertheless, they provide important tests for models and have a modest correlation with long-term climate change in GCMs. Across a broad range of studies, models generally show skill in their ability to reproduce observed humidity change and radiative responses under variability from seasonal to interannual, and under the temperature increase of recent decades.

The sensitivity of feedback strength to base climate is also now much better understood. Paleo reconstructions and modelling studies indicate sensitivity of both water vapour and lapse rate feedback to global temperature and boundary forcing, such as from ice sheets. These, combined with modelling studies with strong warming/cooling from very large changes in CO_2, suggest increasing water vapour feedback with global temperature, including from features such as a heightening tropopause, partially compensated by increasingly negative lapse rate feedback. From a RH based feedback paradigm this implies little change in the Planck or lapse rate feedbacks. Sensitivity of feedbacks to different forcing agents such as solar, black carbon, sulphates, ozone and volcanoes are better understood, with some but not all producing feedback strength of comparable value to CO_2. Under different forcings, as in many other aspects, a high degree of compensation between stronger/weaker water vapour/lapse rate feedback is apparent, again suggesting that the insensitivity to forcing is perhaps better framed in a RH based approach where this offsetting is effectively removed. For very large warming the "partial-Simpsonian" theory predicts total closure of the atmospheric water vapour window as continuum absorption overwhelms other radiative processes (Jeevangee, 2018).

Traditionally defined lapse rate feedback is now much better understood. The overall paradigms of negative feedback in the tropics/subtropics from lapse rate constrained at saturated adiabatic has been consistently reproduced over generations of models, and is also consistent with theoretical understanding and observations of current climate and its interannual variability. Unchallenged demonstration of agreement between model/observations on upper tropospheric temperature *trends* continues to be elusive. Indications of systematic amplified warming in the upper troposphere in models compared to observations appear to be the result of several different factors. First, there are major difficulties in constructing universally accepted long-term observational satellite data sets. Secondly, comparisons can be confounded by natural variability, the incomplete inclusion of forcing such as from volcanoes or and anthropogenic aerosols in model studies or observations, and thirdly from errors in surface temperature in model simulations propagating temperature differences aloft. Although these issues are not fully resolved, atmospheric models forced with observed SSTs show reasonable agreement with observations, and some but not all coupled GCMs show consistency with the somewhat uncertain observations.

The intimate and opposing relationship between water vapour and lapse rate feedbacks has been clarified through theoretical, observational and modelling advances. Globally, the opposing nature results from differences in tropical/extratropical warming, implicating processes such as southern hemisphere sea Ice cover and delayed Southern Ocean warming, which can differ across models. It is now well established that the spread in tropical combined water vapour plus lapse rate feedback results from relative humidity changes, rather than the magnitude of upper tropospheric warming. This places the RH based decomposition of feedbacks on even surer footing, and in recent years increased emphasis has been given to this approach in the literature.

The role of feedbacks in contributing to the amplitude, timing or progression of "modes" of variability such as ENSO, MJO and the ITCZ are also better understood due to observational and modelling studies, but there remains scope for further research in this area to improve understanding of the role of feedbacks in other types of variability.

The importance of feedbacks, particularly lapse rate feedback, in the amplification of polar warming has been appreciated for at least two decades, but with much clarifying research in the last 20 years. Reinforcing interactions between lapse rate feedback, surface albedo feedback, and other feedbacks and processes amplify polar warming, as confirmed by observation and modelling studies. Different studies, however, have found greater or lesser roles for individual feedbacks. Other processes including equator to pole gradients of Planck cooling and CO_2 forcing may also play important roles. Substantial differences also occur between the Arctic and Antarctic, with the latter affected by delayed warming due to ocean heat uptake, and the effect of the elevated Antarctic plateau on lapse rate feedback and other processes. In the absence of an overlying theoretical framework, and in the face of a large number of methodological approaches, the precise quantitative contribution for water vapour and lapse rate feedback in polar amplification remain somewhat elusive, although it is clear that lapse rate feedback plays a strong amplifying role.

Globally, many estimates have been made of the strength of water vapour, lapse rate and the combined feedback. There is strong consistency in the mean and ranges of feedbacks over the last four generations of GCMs, with values also consistent with estimates from observations, such as from trends and from interannual variability scaled by various techniques to quantify long-term climate feedback. The evidence is now overwhelming that combined water vapour+lapse rate feedbacks provide the strongest positive feedback in the climate system, of a magnitude around that produced in climate models. Our estimate of overall strength of these combined feedbacks is $1.25 \pm 0.15\ Wm^{-2}K^{-1}$.

Although, as discussed in the next section, issues remain to be further clarified about water vapour and lapse rate feedbacks, it is extremely unlikely that these will result in major revisions in our confidence or their estimated combined feedback strength.

B. A look to the future: current research gaps

Despite this impressive progress, a range of key issues remain to be fully addressed. Further research is needed to:

1) Improve estimates of water vapour and lapse rate feedbacks from interannual variability. Studies to date vary widely in their conclusions on water vapour feedback strength in particular (Table 1), with differences in approaches, periods, data, and analysis methods behind much of this spread. Refining observational and modelling estimates could help test and verify physical processes in models controlling upper tropospheric temperature and water vapour changes, clarify potential links with feedbacks under secular climate change and hold the hope of improving observational based constraints of feedback strength.
2) Increase understanding of processes underlying the spread in relative humidity changes in the tropics under warming, including separately over land and oceans –

as different responses in models drive much of the uncertainty in the combined water vapour/lapse rate feedback. Uncertainties include factors behind differing patterns of projected warming, and uncertainties introduced by choices in convective parametrisation and other micro-physics choices. Comparing the observed and model-simulated patterns of relative humidity change will be important in this regard. This challenge links closely with the World Climate Research Program's Grand challenge on clouds, circulation and climate sensitivity (Bony et al., 2015).

3) Further develop, understand and apply the RH based approach for decomposing water vapour, lapse rate and Planck feedbacks. Areas include understanding of surface feedbacks and reasons for combined feedback spread (e.g. Po-Chedley et al., 2018; Zelinka et al., 2020; Zhang et al., 2020a).
4) Further explore the promising area of spectral based feedbacks. These have the potential to produce much greater fundamental understanding of changes in water vapour feedback strength with temperature, and indeed of a possible "peak" in overall climate sensitivity as temperatures continue to increase (Seeley and Jeevanjee, 2021; Kluft et al., 2021).
5) Further assess the ability of climate models to simulate the "bimodality" of the water vapour distribution and determine whether the range in skill is of consequence in their representation of feedback processes under both climate variability and change.
6) Determine whether or not convection or cloud microphysics is playing a role in large-scale atmospheric circulation and therefore in controlling the humidity distribution. The very successful advection-condensation, "AC", approach for modelling water vapour distribution does not rule out the possibility of micro-physical induced changes in *broadscale winds* impacting moisture distribution, and therefore water vapour feedback (Dessler and Minschwaner, 2007; Sherwood et al., 2010b).
7) Better understand and model processes involved in aggregation of tropical convection, its response to warming, and the impact on water vapour feedback of this change. If self-aggregation increases, this may weaken water vapour feedback due to the increased areas of tropical and subtropical radiative cooling and changes in high cloud shielding (Wing et al., 2020). Multi-model comparisons to date have found that under SST warming the models were roughly 50:50 split on simulating increased/decreased self-aggregation. The use of a hierarchy of models in this project is a promising direction for understanding processes and sensitivities, but key questions remain on the reasons for model disagreement, and the implications for water vapour feedback and climate sensitivity generally.
8) Better understanding of long-standing apparent disagreements between observed and modelled tropical lapse rate trends over recent decades. This includes better understanding of the differences in the observational datasets.
9) Improve understanding of stratospheric/tropospheric processes, including mechanisms for possible changes in lower stratospheric humidity under global warming. This requires improved observations of changes in stratospheric humidity, better understanding of the effects of ozone and stratospheric water vapour feedback and understanding of how these processes are represented in models. Recent evidence is mixed. Kernels based estimates from CMIP5 models suggest a substantial positive feedback from stratospheric moisture increases, which although substantially weaker than the tropospheric feedback, is nevertheless an important possible contributor to climate sensitivity, of the order of the strength of surface albedo feedback in models (Banerjee et al., 2019). However contrary evidence has been found in a single model from comparison of locked and unlocked stratospheric water vapour experiments, which suggested negligible additional surface warming (Huang et al., 2020).

10) Test suggestions that there may be robust links between the magnitude of water vapour/lapse rate feedbacks and cloud feedback across models (Huybers, 2010). Testing of such links and, where robust, improved understanding of processes could shed light both on cloud feedbacks and water vapour/lapse rate feedbacks.
11) To explore beyond the linear feedback assumptions that are prevalent in much of the literature (Lahellec et al., 2008; Knutti and Rugenstein, 2015). There is ample evidence of nonlinear evolution of feedbacks with warming and forcing, and understanding and quantifying these are important for increased confidence in future climate response (Knutti and Rugenstein, 2015). The issue becomes steadily more important as global warming progresses later this century and beyond, and full use could be made of co-ordinated intercomparisons of very long timescale, strongly forced scenarios (Rugenstein et al., 2019).
12) Further clarify the role of feedbacks in high latitude amplification of warming. Although it is clear that lapse rate feedback is important, interactions are complex, with feedbacks operating at the TOA, within the atmospheric column and at the surface. Some contributions, such as stratospheric water vapour have been little considered (Li and Newman, 2020). An overall unifying theory of the key processes would shed much light on polar amplification and the role of radiative feedbacks. As Russotto and Biasutti (2020) aptly point out "A multi-GCM study perturbing all relevant feedbacks … might help to resolve the disagreements over the causes of polar amplification obtained from limited GCM experiments and different diagnostic techniques."

Very appropriately, the focus of the research community in climate feedbacks over the last 10-15 years has moved to better understanding and constraining cloud feedback (Bony et al., 2015; Sherwood et al, 2020). This is due to an appreciation that differences in cloud feedbacks across models are very large, and responsible for much of the spread in resulting climate sensitivity (Bony et al., 2006; 2015). Nevertheless, significant issues remain unresolved in understanding and modelling water vapour and lapse rate feedbacks, and inter model spread in the combined feedback is the second largest source of uncertainty in determining the value of the ECS (DuFresne and Bony, 2008). Given the magnitude of water vapour and lapse rate feedbacks, and their fundamental role in projected climate change it is imperative that they receive appropriate focus in upcoming years.

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I. Appendix 1: Quantifying water vapour and lapse rate feedbacks in models and observations

Major advances in quantifying feedbacks have been a key factor in better understanding of feedbacks over the last three decades. These fall into several categories:

1. TOA clear sky radiation changes

The simplest method, and one long used, evaluates "clear sky" changes under warming from the instantaneous zeroing of clouds prior to radiation calculations (e.g. Webb et al., 2006). In the LW, this methodology provides a convolution of all temperature and LW water vapour responses (including the "Planck" response, water vapour and lapse rate) and removes the (sometimes very large) impact of cloud cover on these changes (Soden et al., 2004). In the SW it is a convolution of SW water vapour impacts with surface albedo changes, again with the effect of clouds removed. Hence, these are not the radiation changes from the "true" feedbacks "seen" in the real world or by a GCM. The methodology can, however, be useful where limited fields are available from models or many models are being compared (Andrews et al, 2015) and more sophisticated approaches cannot be used. Most model simulations routinely archive the results of a "cloud radiative effect" calculation, whereby the radiation code is run once with "all sky" conditions, then a second time with clouds removed, so the required analysis fields are very widely available.

2. Partial radiative perturbation

A second, and much more accurate, approach – the "partial radiative perturbation" method (PRP) evaluates the radiative impact of feedbacks directly by performing instantaneous (e.g. daily) calculations within GCM radiation code, with water vapour and temperature changes swapped one by one between the climates under examination—e.g. current climate and future warmed climate (Wetherald and Manabe, 1988). Care must be taken because the field swapping introduces a radiation bias because of de-correlations between the variables (Colman and McAvaney, 1997; Schneider et al., 1999; Klocke et al., 2013), necessitating a second "reverse swap" from the current climate into the future, then differencing of the results to remove this bias (Soden et al, 2008). The PRP approach, although having the advantage of high accuracy and clear separation of feedback variables, has the downside of being extremely computationally and logistically intensive. There is also significant interannual variability of the measured strength of these feedbacks, with one study finding differences of 0.5-1.0 $Wm^{-2}K^{-1}$, implying averaging periods of 3 and 5 years for accurate estimates of lapse rate and water vapour feedbacks respectively (Klocke et al., 2013). PRP has been used extensively for studies such as response to different forcings (Yoshimori and Broccoli, 2008), climate change under forcing scenarios (Colman et al., 2001), and paleo experiments (Yoshimori et al., 2009).

An extension beyond the PRP approach (the so-called CFRAM method) diagnoses all fluxes contributing to each of the traditional feedbacks (including water vapour and lapse rate), as represented by partial temperature change contributions at each point in latitude longitude and height including non-radiative processes such as dynamical changes and surface fluxes (Lu and Cai, 2009; Taylor et al., 2013). Although this analysis differs from traditional feedback approaches, it has the advantage of providing well-defined fractional influences on temperature change from physical processes at each point in time and space, such as surface radiative contributions to amplified high latitude warming (Sejas and Cai, 2016).

3. Radiative kernels

A third approach which has become a standard methodology over the last decade is the use of "radiative kernels" (Soden and Held, 2006; Soden et al., 2008; Shell et al., 2008). This approach divides the total radiative response $\lambda_x = \frac{\partial R}{\partial x}\frac{\partial x}{\partial T_s}$ into two terms: radiative transfer and climate response. The radiative transfer term is derived from one-sided PRP type calculations within a single model employing standardised perturbations on top of its base climate. Relevant to the evaluation of lapse rate and water vapour feedbacks, are kernels derived from +1K temperature increases with unchanged specific humidity, and with fixed relative humidity. These are then applied to other GCMs, by multiplying the kernel by the temperature changes a function of height latitude and month found in that GCM (the climate response). Commonly, kernels may be produced for both clear sky and all sky conditions, so is to study the impact of clouds on individual feedbacks. Details of the methodology are provided in Soden and Held (2006). This approach provides close approximations of the PRP methodology (Soden and Held, 2006) and permits wide comparison of feedback strengths in GCMs including from multiple experiments such as perturbed physics ensembles (Shell et al, 2008; Sanderson et al, 2010). An important feature of radiative kernels is they can also be used on observational or reanalysis data to provide estimates of radiative impacts from climate variability or change (Dessler, 2013; Colman and Hanson, 2012). The use of monthly means as field input, and the requirement for only a single "forward" calculation, make this a very attractive alternative to PRP. A substantial number of kernels have been derived from different GCMs (Soden et al., 2008; Shell et al., 2008; Block and Mauritsen, 2013; Pendergrass et al., 2018; Smith et al., 2018) and made widely available for research applications.

Due to the state dependency of the kernels, if the climate moves far from the present (e.g. under much stronger CO_2 forcing, such as $4xCO_2$ or above) then the methodology leads to inaccuracies, necessitating recalculation of the kernels at a more appropriate, e.g. warmer, climate (Jonko et al., 2012; Ceppi and Gregory, 2017).

An important approximation inherent in the methodology is that the kernels are calculated under the "climate" of one particular model and using a separate radiation scheme to that of the GCM(s) being studied. Different base model temperature, water vapour and cloud climatologies between kernel and target GCM can result in in different diagnosed radiative impact from temperature or water vapour changes to those "seen" in the original target GCM climate experiment (Soden et al., 2008). A number of studies have found these effects to be relatively small however (e.g. Soden et al., 2008), and a recent study applying radiative kernels derived from six different GCMs found these effects to be overall unimportant in evaluating and comparing quantities such as global mean water vapour and lapse rate feedback (Zelinka et al., 2020). It has been found that relatively high vertical resolution of the stratosphere may be needed to resolve temperature lapse rate or water vapour feedbacks in this region (Smith et al., 2020). Another issue when using kernels to estimate stratospheric changes in temperature and moisture, is the need for the kernel to take into consideration rapid tropospheric temperature adjustments, as radiative affects are sensitive to these temperature changes (Maycock et al., 2011; Banerjee et al., 2019).

4. Cutting feedback loops

The three methods above essentially represent "postprocessing" of GCM or observed results. A final approach has been to "cut the feedback loop" i.e. isolate and suppress the radiative response from changes in water vapour and/or lapse rate in climate model experiments (Schneider et al.; 1999; Hall and Manabe, 1999, 2000a,b; Langen et al., 2012; Mauritsen et al., 2013; Henry and Merlis, 2020; Byrne and Zanna, 2020). Comparing such "decoupled" experiments with standard model runs permits examination of the effect on associated physical processes and their response to the warming/cooling associated with the feedback, and has been shown to provide an extremely clean separation of feedbacks, and close agreement with PRP approaches. There needs, however, to be careful treatment of de-correlation issues between fields when calculating radiation, which affects both the unforced climate and the climate change in response to forcing (Mauritsen et al., 2013). Using this approach Hall and Manabe (1999, 2000a, b) demonstrated directly that water vapour feedback not only amplifies climate change, but also unforced "natural" variability in a coupled GCM. The method has also been useful for comparisons with observations in response to volcanic forcing (Soden et al., 2002). This approach has also been used in a range of experiments examining the causes of high latitude amplification, by systematically suppressing one or more feedback processes (e.g. Langen et al., 2012; Henry and Merlis, 2020), the role of water vapour feedback in the seasonal shift in the ITCZ (Clark et al., 2018), and the role of water vapour feedback in understanding the seasonal progression of the monsoon, and its response to climate change (Byrne and Zanna, 2020).

II. Appendix 2. Evaluations of water vapour and lapse rate feedbacks

Reference	Dataset(s)	Method	Value ($Wm^{-2}K^{-1}$)
Cubasch and Cess (1990)	Cess et al. (1989) Raval and Ramanathan (1989).	Assessment from literature	1.2 (Net)
Colman (2001)	RCMs and GCMs	Reported range in literature	1.7 ± 0.78 (Net) -0.32 ± 0.78 (LR) 1.37 ± 0.4 (WV+LR)
Forster and Collins (2004)	Post Pinatubo cooling. NASA Water Vapor Project (NVAP), Microwave Limb Sounder (MLS)	Off-line radiation calculations on satellite moisture retrievals	1.6 (0.9-2.5) (Net)
Soden and Held (2006)	CMIP2 models, climate projections	Kernels applied to CMIP2	1.81 ± 0.53 (Net) -0.87 ± 0.72 (LR)
Soden and Held (2006)	CMIP3 models, climate projections	Kernels applied to CMIP3 SRES A1B	1.80 ± 0.36 (Net) -0.84 ± 0.52 (LR)
Gordon et al. (2013)	AIRS sounder 2002-2009	Off-line radiative transfer model with observed water vapour distribution and CMIP3 model long-term/variability ratio	1.9-2.8 (Net)
Dalton and Shell (2013)	CMIP3 Models over historical warming	Kernels	1.79 ± 0.26 (LW)
Dalton and Shell (2013)	ERA-Interim (1989-2008)	Monthly variability scaled by model long-term/interannual	1.67 (0.48-1.91) (LW)
Koumoutsaris (2013)	CMIP3 models, climate projections	Kernels	1.88 ± 0.26 (Net) -0.84 ± 0.36 (LR) 1.2 ± 0.24 (WV+LR)
Vial et al. 2013	CMIP5 4xCO_2	Kernels (average of 2 used)	1.65 ± 0.28 (Net) -0.60 ± 0.40 (LR) 1.05 ± 0.11 (WV+LR)
Boucher et al. (2013)	Multiple, including CMIP5 projections	Models and assessment of broad evidence	1.1 (0.92 to 1.3) (90% range)
Chung et al. (2014)	CMIP5 "Historical" experiments	Kernels applied to warming between 1979-1988 and 1989-1998.	1.92 ± 0.99 (Net).
Caldwell et al. (2016)	CMIP5 4xCO_2	Kernels	1.70 ± 0.36 (Net) -0.55 ± 0.50 (LR)
Colman and Hanson (2017)	CMIP5 RCP 8.5	Kernels	1.75 ± 0.38 (LW) 2.01 ± 0.40 (Net) -0.74 ± 0.46 (LR)
Colman and Hanson (2017)	CMIP5 4xCO_2	Kernels	1.58 ± 0.40 (LW) 1.82 ± 0.41 (Net)

			-0.42 ± 0.56 (LR)
Liu et al. (2018)	AIRS-MLS	Interannual observed value scaled by model long-term/variability ratio	1.85 ± 0.32 (Net)
Zelinka et al. (2020)	CMIP5 4xCO_2	kernels	1.81 ± 0.38 (Net) -0.48 ± 0.48 (LR)
Zelinka et al. (2020)	CMIP6 4xCO_2	kernels	1.83 ± 0.28 (Net) -0.50 ± 0.36 (LR)
Sherwood et al. (2020)		Assessment from literature	1.15 ± 0.30 (WV+LR)

Table A2-1. Summary of estimates of water vapour and lapse rate feedbacks from models and observations. SRES is the Special Report on Emissions Scenarios (Nakicenovic et al., 2000). RCP8.5 is radiative concentration pathways scenario 8.5 (Van Vuuren et al., 2011). Range of values shown is $\pm 2\sigma$. LW is the long wave component of the water vapour feedback only. Net is LW+SW. LR is lapse rate feedback.

III. Appendix 3: Acronyms

Acronym	Meaning
AC	Advection-condensation
AIRS-MLS	Atmospheric Infrared Sounder-Microwave Limb Sounder
CERES	Clouds and the Earth's Radiant Energy System
CMIP	Coupled Model Intercomparison Project
CRM	Cloud resolving model
ECMWF	European Centre for medium-range weather forecasting
ECS	Equilibrium climate sensitivity
ENSO	El Niño Southern Oscillation
ERA	ECMWF reanalysis
GCM	Global climate model
GHG	Greenhouse gas
GFDL	Geophysical Fluid Dynamics Laboratory
GPS	Global positioning system
IPCC	Intergovernmental Panel on Climate Change
ITCZ	Inter-tropical Convergence Zone
JRA-25	Japanese 25-year Reanalysis
LGM	Last glacial maximum
LW	Long wave (terrestrial) radiation
MJO	Madden-Julian Oscillation
MERRA	Modern Era Retrospective-Analysis for Research and Applications
MSU/AMSU	Microwave sounding unit/Advanced microwave sounding unit
OLR	Outgoing long wave radiation
PRP	Partial radiative perturbation
RCM	Radiative-convective model
SALR	Saturated adiabatic lapse rate
SST	Sea surface temperature
SW	Short wave (solar) radiation
TOA	Top of atmosphere
TOVS	Television Infrared Observation Satellite (TIROS) Operational vertical sounder

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