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Quantifying the water vapour feedback associated with post-Pinatubo global cooling

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Abstract There is an ongoing important debate about the role of water vapour in climate change. Predictions of future climate change depend strongly on the magnitude of the water vapour feedback and until now models have almost exclusively been relied upon to quantify this feedback. In this work we employ observations of water vapour changes, together with detailed radiative calculations to estimate the water vapour feedback for the case of the Mt. Pinatubo eruption. We then compare our observed estimate with that calculated from a relatively large ensemble of simulations from a complex coupled climate model. We calculate an observed water vapour feedback parameter of –1.6 Wm^{-2} K^{-1}, with uncertainty placing the feedback parameter between –0.9 to –2.5 Wm^{-2} K^{-1}. The uncertain is principally from natural climate variations that contaminate the volcanic cooling. The observed estimates are consistent with that found in the climate model, with the ensemble average model feedback parameter being –2.0 Wm^{-2} K^{-1}, with a 5–95% range of –0.4 to –3.6 Wm^{-2} K^{-1} (as in the case of the observations, the spread is due to an inability to separate the forced response from natural variability). However, in both the upper troposphere and Southern Hemisphere the observed model water vapour response differs markedly from the observations. The observed range represents a 40%–400% increase in the magnitude of surface temperature change when compared to a fixed water vapour response and is in good agreement with values found in other studies. Variability, both in the observed value and in the climate model's feedback parameter, between different ensemble members, suggests that the long-term water vapour feedback associated with global climate change could still be a factor of 2 or 3 different than the mean observed value found here and the model water vapour feedback could be quite different from this value; although a small water vapour feedback appears unlikely. We also discuss where in the atmosphere water vapour changes have their largest effect on surface climate.

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1 Introduction

One of the major uncertainties in our ability to predict future climate change is knowledge of the magnitude of climate response associated with an initial climate forcing – the climate feedback. IPCC (1990) suggested a range of 1.5–4.5 K for the global surface temperature increase associated with a doubling of CO_2. Since then, despite a massive improvement in models and in our understanding of the mechanisms of climate change, the uncertainty in our projections of temperature change has stubbornly refused to narrow: IPCC (2002), gives an identical range.

Whilst uncertainties in the cloud response (e.g. Cess et al. 1996) probably contribute most to this uncertainty, as the water vapour feedback is expected to be large, even a small uncertainty in it can have a large effect on climate predictions. Attempts to measure the water vapour feedback from observations of regional, seasonal and interannual changes in water vapour have been difficult to interpret (Bony et al. 1995; Lau et al. 1996;

Inamdar and Ramanathen 1998). Accurate quantification relies on an understanding of the mechanisms involved and these are still debated (Lindzen 1990; IPCC 2002). However, in spite of these drawbacks it is generally believed that the water vapour feedback is both large and positive, roughly doubling the expected surface temperature response (IPCC 2002).

In an important paper, Soden et al. (2002), for the first time, examined the observed water vapour response to a global climate change (the eruption of Mt. Pinatubo). They compared observations with three ensemble pairs of a general circulation model. They found that timeseries of globally averaged total column water vapour (1000 hPa–300 hPa) and globally averaged column upper tropospheric water vapour (500 hPa–300 hPa) were well simulated by the model, and with a version of their model that excluded water vapour changes in the radiative calculation there was an underestimation of the associated tropospheric temperature changes, whereas the full model was able to reproduce these temperature changes.

Their work provided circumstantial evidence for similar water vapour feedbacks in the observations and model. However, as the authors themselves acknowledge, their findings are unable to completely confirm the relationship. Firstly, the large positive water vapour feedback found in their model could have, in the observations, come from other sources, such as cloud changes. Secondly, due to the dependence of the radiative forcing of water vapour changes with latitude and height (discussed in Sect. 3) it is possible for similar globally averaged total column water vapour changes to have very different effects on the Earth's energy budget and lead to quite different water vapour feedbacks.

Using their paper as inspiration, we go beyond examining globally averaged column water vapour changes. We use measurements of the *pattern* of water vapour changes to *quantify* the water vapour feedback in both the observations and a different climate model.

2 Data and Methodology

2.1 Data

Surface temperature data used in this study were monthly averaged anomalies taken from the Climate Research Unit (CRU) data set (Jones et al. 1999). As in Soden et al. (2002), water vapour data was obtained from the NASA Water Vapour Project (NVAP) (Randel et al. 1996). Monthly data of precipitable water for three layers (surface-700 hPa, 700–500 hPa and 500–300 hPa) on a $1° \times 1°$ resolution grid were used covering the period 1988–1997. To extend our water vapour analysis above 300 hPa we also employed data from the Microwave Limb Sounder (MLS) instrument. This instrument gauges thermal emission at 1.5 mm and is not contaminated by the presence of a stratospheric aerosol layer. The dataset was an improved version of the dataset described by Read et al. (1995) (Read personal communication). Data was obtained for ppmv of water vapour in 3 km thick layers centred on 147 hPa and 215 hPa, approximately covering the 300–100 hPa region. Individual soundings were binned into monthly averages for six latitude bands (0°–30°, 30°–60°, 60°–90°, in each hemisphere). Data was only available between September 1991–June 1997. Where necessary the average of the same month from the complete timeseries was used as an estimate of a pre-September 1991 value. Uncertainties in MLS retrievals are estimated to be 20% (Read et al. 1995).

The model used is version three of the Hadley Centre for Climate Prediction and Research coupled atmosphere ocean model HadCM3 (Gordon et al. 2000; Collins et al. 2001). The ocean component of the model has a horizontal resolution of 1.25° longitude by 1.25° latitude and has 20 levels in the vertical from the surface to the bottom of the ocean. The atmospheric component of the model has a horizontal resolution of $3.75° \times 2.5°$ in longitude and latitude with 19 unequally spaced vertical levels from the surface up to approximately 40 km, with six levels representing the stratosphere. The two components are coupled without the use of artificial flux corrections and the model has a stable surface climate. Aerosol distributions associated with Pinatubo are taken from an updated version of the Sato et al. (1993) dataset and are introduced into the model in four uniform latitudinal bands from 90°S–45°S, 45°S-equator, equator-45°N and 45°N–90°N. An ensemble of 13 simulations are analysed, each starting in December 1985 with a variety of different atmospheric and oceanic initial conditions. Collins (2003) gives a detailed analysis of the surface climate response of the model to the eruptions of both Pinatubo and El Chichón.

2.2 Radiative transfer calculations

Once the water vapour changes have been analysed we need to calculate their effects on the Earth's radiation budget and a water vapour climate feedback parameter.

The global mean equilibrium surface temperature change (ΔT_S) can be written as

$$\Delta T_s = \Delta F / Y \tag{1}$$

where ΔF is a radiative forcing and Y is the climate feedback parameter. If feedbacks are ignored Y is chosen so Eq. 1 gives the surface temperature change that would result if the Earth behaved as a blackbody ($Y = Y_{black\_body}$ ≈3.3 Wm^{-2} K^{-1}) (see e.g. Cess et al. 1990). Assuming a linear feedback model, other feedbacks can then be included in the system

$$Y = Y_{black\_boby} + Y_{water\_vapour} + Y_{cloud} + Y_{surface\_albedo} \cdots \tag{2}$$

The water vapour feedback, for example, would be written as

$$Y_{water\_vapour} = -\partial Q/\partial q . \partial q/\partial T_s \approx -\Delta Q_{water\_vapour}/\Delta T_s \tag{3}$$

where $\partial Q/\partial q$ gives the change in the Earth's energy balance for a given change in water vapour and $\partial q/\partial T_s$ gives the change in water vapour for a given change in global mean surface temperature. From Eq. 3 it follows that the water vapour feedback parameter (Y_{water_vapour}) can be calculated as the change in Earth's energy balance from water vapour changes per unit change in global mean surface temperature.

In terms of estimating a Y_{water_vapour} which is relevant to long-term climate change, we calculated the annually averaged perturbation to the Earth's energy balance from water vapour changes which were imposed on a long-term climatology, rather than calculate the transient radiative effects of water vapour in the particular the case of Pinatubo. We first carried out a set of sensitivity experiments that were subsequently used to weight observed water vapour changes to calculate a change in the Earth's energy balance, used in Eq. 3.

1. We employ the Reading Narrow Band Model (NBM) in the longwave (Shine 1991), a high resolution shortwave radiation scheme (Forster and Shine 1997) and a seasonal zonally averaged 5° latitudinal resolution climatology of water vapour, ozone, temperature and clouds, based on European Centre analysis data and International Satellite Cloud Climatology Project data (see Forster and Shine 1997). In this model we increased the water vapour at different latitudes and heights by a constant 10% and calculated the perturbation to the Earth's energy balance from these changes. In contrast to previous studies, rather than simply calculate the change in outgoing longwave radiation (OLR) our calculations included the effects of water vapour in the shortwave and a stratospheric adjustment term (see e.g. Forster and Shine 1997). These improvements to the calculation were found to increase the net effects of water vapour on the energy budget by approximately 10%. The radiative perturbations calculated from this sensitivity study are shown in Fig. 1.
2. The radiative perturbations from Fig. 1 were subsequently used as a look up table to weight the observed and model water vapour changes to calculate the ΔQ term of Eq. 3. Whilst this methodology neglects the effects of any non-linearity in the radiative calculation, tests showed that this non-linearity modified the calculate forcing by no more than 10% for the relatively small water vapour changes associated with Pinatubo. This error was insignificant compared to the errors in the water vapour changes themselves.

Using an estimate of the global mean surface temperature change from either CRU data (for the observations) or HadCM3 data (for the model) an estimate of the water vapour feedback parameter was obtained from Eq. 3. Eq. 3 is derived for equilibrium studies; however, it has also been shown to be a reasonably representative

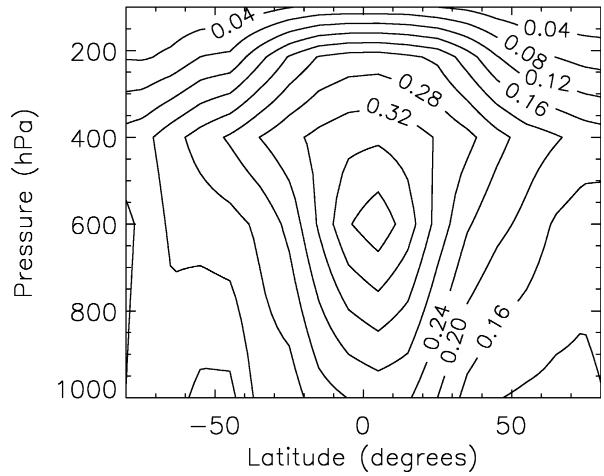


Fig. 1 The change in Earth's energy balance (Wm^{-2}) for a 10% percentage change in water vapour within a 100 hPa slab of atmosphere centred on the y axis pressure value

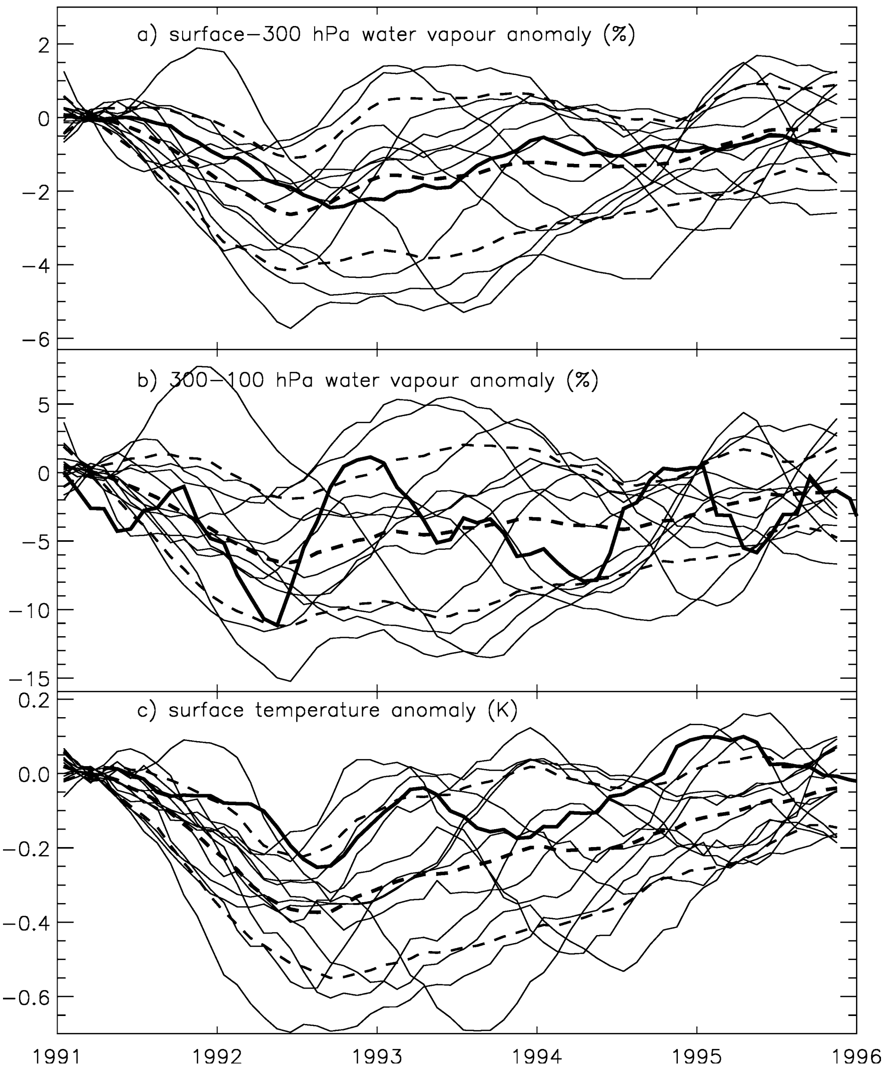
of transient climate changes, provided enough time has elapsed for the feedback mechanism to respond to the surface temperature change (e.g. Hansen et al. 1993). We assume that this time scale is less than 30 days for tropospheric water vapour.

3 Results and discussion

Time series of globally average water vapour and surface temperature anomalies are shown for the observations and HadCM3 model ensemble in Fig. 2. As noted by Soden et al. (2002), there is a well-modelled 2% decrease in column water vapour below 300 hPa after the eruption (Fig. 2a). We have used many more ensemble members than in the Soden et al. study and although a significant drying is found in most members, it is not a ubiquitous feature. Climate "noise" can in some cases mask the volcanic signal. Figure 2 also shows water vapour changes in the 300–100 hPa region (Fig. 2b), not examined in Soden et al. (2002), and the surface temperature changes (Fig. 2c). Although the MLS vapour time series is more variable than the tropospheric water vapour, it also suggests up to a 10% drying in mid 1992, 1 year after the eruption. The CRU temperature observations show a 0.2 K surface cooling towards the end of 1992. Many ensemble members capture both this drying and the surface cooling.

Using the time period of observed maximum surface temperature response (April–December 1992), Fig. 3 shows the zonal mean patterns of water vapour change from pre-Pinatubo values. The observations show a definite hemispheric difference in response especially in the upper troposphere, with water vapour increases in the Southern Hemisphere and decreases in the Northern Hemisphere (both the MLS and NVAP data show this). The percentage water vapour change also generally

Fig. 2 Time series of **a** surface-300 hPa column water vapour anomaly (%), **b** 300–100 hPa column water vapour anomaly (%) and **c** surface temperature anomaly (K). As in Soden et al. (2002), the seasonal cycle has been removed and anomalies are taken using, as a base, the average values during the five-months prior to the Pinatubo eruption (Jan–May 1991); a 7-month running mean has also been applied to the data. *Solid lines* are from observations and *thin grey lines* are 13 individual ensemble members. The *thick dashed line* shows the average of all ensembles and the *thin dashed lines* represent the +1 and –1 standard deviation from the ensembles



increases with height. Nearly all the model ensemble members have a similar increase in water vapour change with height, however none capture the hemispheric difference. The wide range of patterns found between ensemble members, especially near the equator, illustrates the large variability in the model water vapour response; although when the time period was varied by up to four months, the broader features, such as the general reduction of water vapour in the model, the hemispheric difference in the observations and the increase in water vapour changes with height remained robust.

Figure 4 illustrates where in the atmosphere the water vapour changes most affect the Earth's energy balance (ΔQ). The observations (Fig. 4a) are compared to a model ensemble member (which we denote *e7)* that behaves most like the observations (Fig. 4b). While the magnitude of the energy balance changes in the model and observations are similar, in the observations the radiative influence of water vapour predominately arises from the large MLS trends between 300–100 hPa while in the model ensemble member the radiative influence is largely felt in the mid- tropical troposphere. In the past there has been considerable discussion about where water vapour changes have their largest impact (e.g. Lindzen 1990; Shine and Sinha 1992; Schneider et al. 1999; Allan et al. 1999). In terms of a percentage change in water vapour (Fig. 1) it is the tropical mid troposphere that dominates (see also Shine and Sinha 1992; Colman 2002). However, for a constant relative humidity change the level of importance is higher for two reasons: Firstly, the saturation mixing ratio increases in proportion to T^{-2}; hence this gives roughly twice the percentage change in specific humidity near the tropopause, compared to the surface. Further, in models at least, the lapse typically changes in the tropics, so the temperature change at the tropopause is larger than at the surface. Colman (2002) includes both these effects and finds the most important layer to be 200–400 hPa. Held and Soden (2000) include the temperature dependence of the saturation mixing ratio and find the 300–400 hPa layer to be the most important. Here, for the

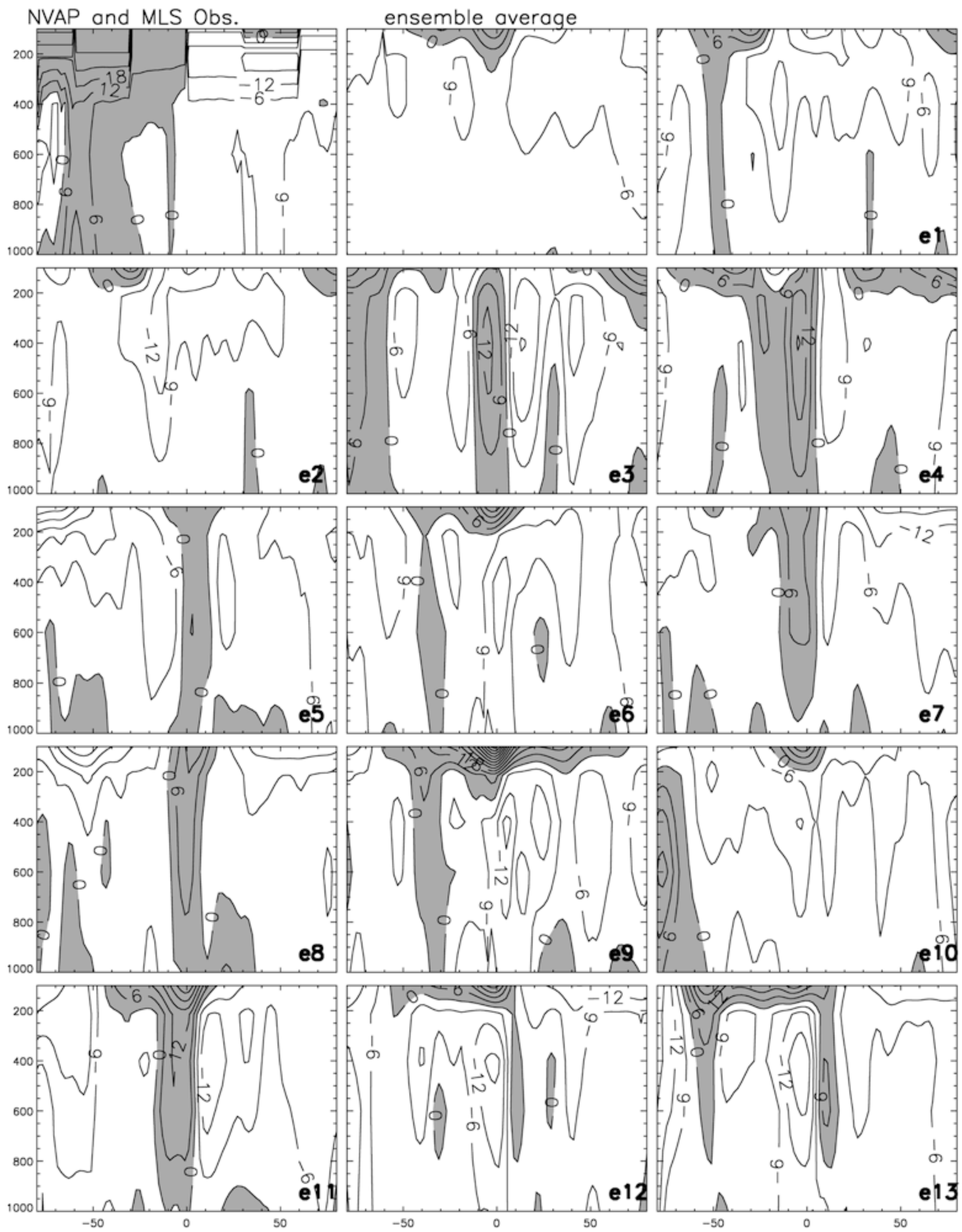


Fig. 3 Water vapour changes in (%) as a function of latitude and pressure. Data is shown for the observations (*upper left panel*), ensemble average (*upper right panel*) and the individual ensemble members (denoted *e1...e13*). The contour interval is 6% and shaded areas represent positive changes. Changes are taken by comparing data for April–December 1992 with the 5-month pre-Pinatubo average value, having firstly removed the seasonal cycle

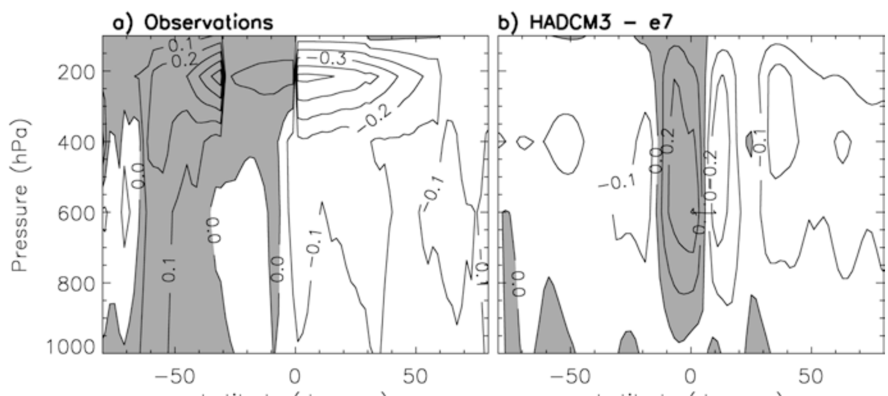


Fig. 4 The change in Earth's energy balance (Wm^{-2}) arising from a 100 hPa thick atmospheric slab centred on the *y*-axis pressure value, for **a** the observed water vapour change and **b** the *e7*ensemble member change. Changes are taken by comparing data for April–December 1992 with the 5-month pre-Pinatubo average value, having firstly removed the seasonal cycle

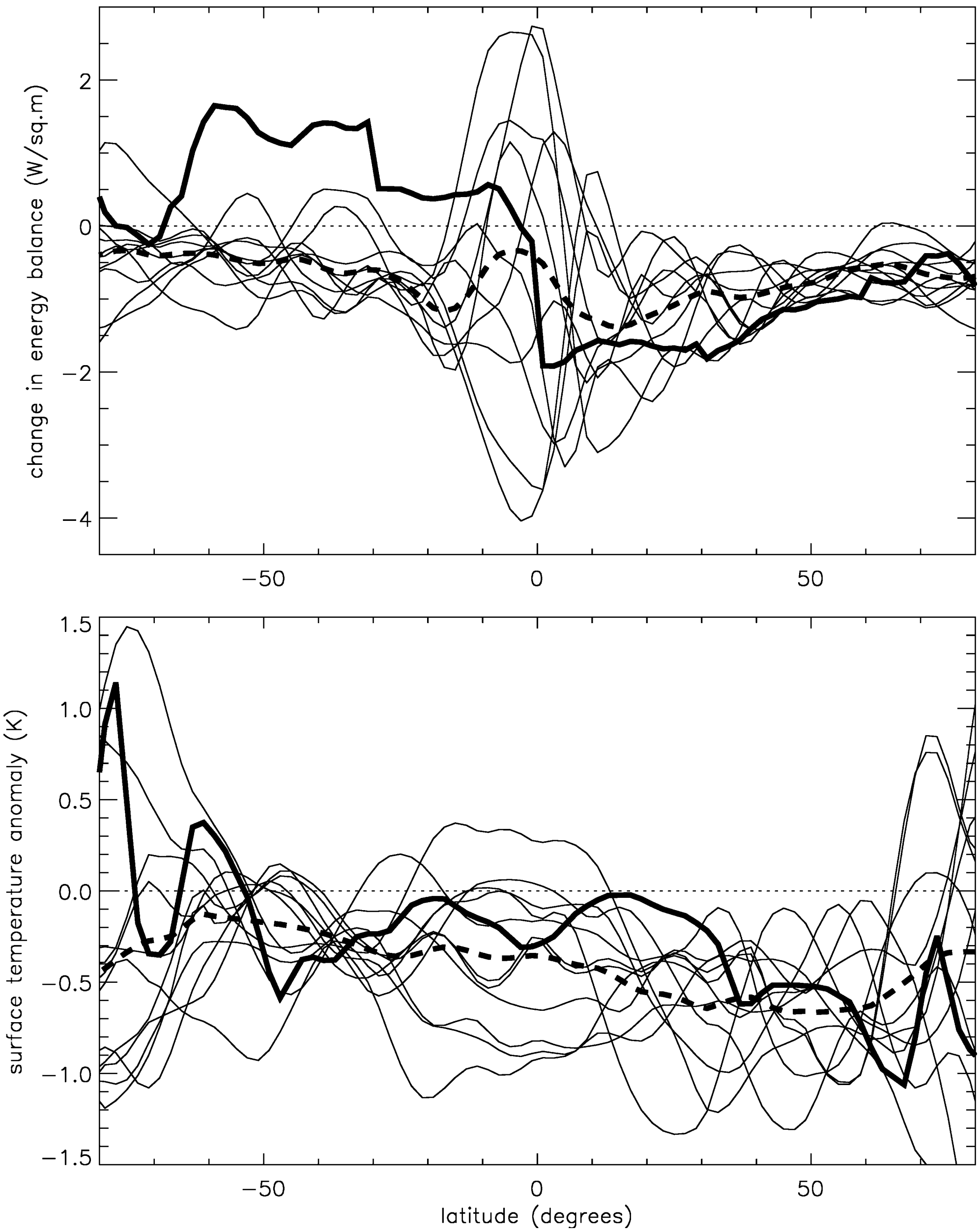


Fig. 5 a The change in Earth's energy balance and **b** the change in surface temperature as a function of latitude. Observations are the *thick solid lines*, the ensemble members the *thin grey lines* and the ensemble averages the *thick dashed lines*. Changes are taken by comparing data for April–December 1992 with the 5-month pre-Pinatubo average value, having firstly removed the seasonal cycle

observations of Pinatubo (Fig. 4a) the upper tropospheric response dominates the water vapour feedback, which is in agreement with these previous findings. However, perhaps due to relative humidity changes, the model response (Fig. 4b) is at odds with these findings from other models.

Experiment *e7* was chosen to compare with observed changes, as the simulated global mean cooling following the eruption most resembled that observed in both timing and amplitude. The model used here has a relatively realistic representation of ENSO variability (Collins et al. 2001) that correlates strongly with global mean temperature. Further, the *e7* ENSO phase is similar to the rather persistent El Nino observed during the period and therefore it is the most appropriate ensemble member for comparison. The ensemble mean is less appropriate as ensemble averaging would reduce the "climate noise" that the observed response retains. When the fields in Fig. 4 are integrated to calculate a globally averaged change in the Earth's energy balance they give similar numbers: –0.33 Wm^{-2} and –0.44 Wm^{-2} for the observations and *e7* respectively.

Other ensemble members have a more negative ΔQ than the observations due to their water vapour decreases in the Southern Hemisphere. The hemispheric asymmetry in the observed energy balance change was not seen in any ensemble member (Fig. 5a), despite several having a similarly persistent El Nino. Forster et al. (2000) found that a forcing confined to one hemisphere generated a larger equilibrium response in that hemisphere, although there was considerable cross-equatorial spreading out of the response. The hemispheric difference in the change of energy balance (Fig. 5a) could be contributing to the observed asymmetry in the surface cooling (Fig. 5b). However, this may be unlikely, as it appears that more ensemble members capture the asymmetry in surface temperature than they do in the energy balance change (Fig. 5a). This could mean that the asymmetry in surface temperature is largely a direct response to the volcanic forcing and due to the transient nature of the Pinatubo eruption not enough time has elapsed for the water vapour feedback to operate.

We employ Eq. 3 with monthly mean estimates of ΔT_S and ΔQ from both the observations and the model ensemble to calculate distributions of Y_{water_vapour}. For the observations, the nine months (April–December 1992) with the largest temperature signal were chosen in order to produce the most robust estimates. For the model ensemble, the same period was chosen and months with $\Delta T_S < 0.1$ K were excluded from the analysis for robustness. This resulted in a total of 82 months from which to compute the histogram of Y_{water_vapour} shown in Fig. 6 together with the observed estimates.

We calculate a mean Y_{water_vapour} of –1.6 $\mathrm{Wm^{-2}\,K^{-1}}$ for the observations and 2.0 $\mathrm{Wm^{-2}\,K^{-1}}$ for the mean of the 82 model months. The range of Y_{water_vapour} from the nine months of observations is –0.9 to –2.5 $\mathrm{Wm^{-2}\,K^{-1}}$ and the 5–95% range from the fitted normal distribution of the model values is –0.4 to –3.6 $\mathrm{Wm^{-2}\,K^{-1}}$. The raw histogram of model values gives a median value of

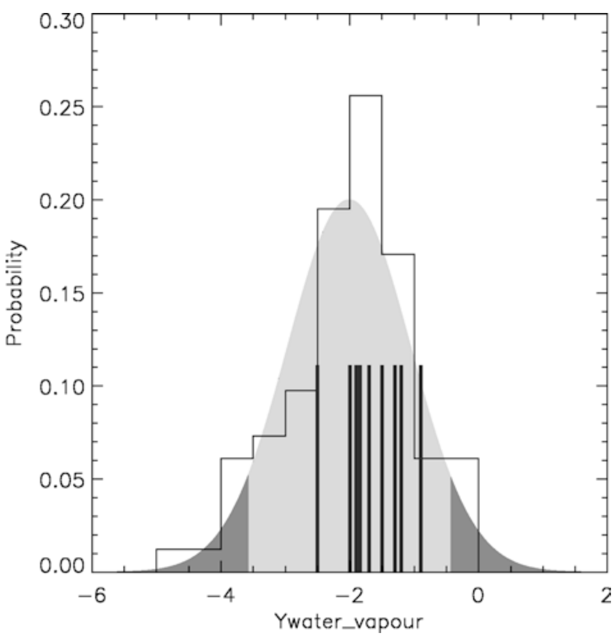


Fig. 6 Estimates of Y_{water_vapour}($\mathrm{Wm^{-2}\,K^{-1}}$) from the observations and from HadCM3. The histogram is computed from 82 model estimates with a bin size of 0.5 and is shown in terms of probabilities. The *shaded curve* is a fitted normal distribution of model estimates with the 5% and 95% represented by *darker shading*. Observed estimates of Y_{water_vapour} are indicated by the *vertical lines*

Y_{water_vapour} –1.5 to –2.0 $\mathrm{Wm^{-2}\,K^{-1}}$ with a probability of 0.3. The uncertainty in the calculation of Y_{water_vapour}, is caused by the inherent variability of the climate system, and is much larger than other sources of error either in the observations or radiative transfer calculations. Nevertheless, despite the large range, it is still encouraging that observed Y_{water_vapour} is consistent with that calculated by HadCM3. It represents roughly a doubling of the surface temperature change you would expect with fixed water vapour and agrees well with estimates from previous model studies (Cess et al. 1990; Hall and Manabe 1999; Schneider et al. 1999; Held and Soden 2000). The variability of Y_{water_vapour} is less than the variability in the surface temperature changes between ensemble members. This suggests that water vapour response differences are not the major source of this variability.

It should be noted that we have essentially calculated a water vapour feedback term for a transient change in stratospheric aerosol. It is now acknowledged that different mechanisms of climate change can have different magnitudes of feedback (Hansen et al. 1997; Joshi et al. 2002). It is therefore possible that a volcanic eruption would have a different water vapour feedback than long-term greenhouse gas increases. For example it is easy to envisage that due to the nature of the forcings, the upper tropospheric responses could be different. Hansen et al. (1997) found that a stratospheric aerosol perturbation, with an absorbing aerosol, had a more negative feedback parameter than a $2 \times CO_2$experiment. In contrast, with a purely scattering aerosol, which would be more representative of a typical volcanic eruption, the greenhouse gas and aerosol feedbacks differed by only 5%. However, this modelling study is a single model result, performed equilibrium experiments and did not explicitly resolve the water vapour feedback.

4 Conclusions

We have, we believe, for the first time attempted to quantify the water vapour feedback for a global climate change: in this case the eruption of Mt. Pinatubo. We find a Y_{water_vapour} of –1.6 $\mathrm{Wm^{-2}\,K^{-1}}$ with a range of –0.9 to –2.5 $\mathrm{Wm^{-2}\,K^{-1}}$. This represents a 40–400% increase of the surface temperature change that you would expect from a simple blackbody response to a forcing.

Differences were found between the model and observed water vapour changes, most noticeable were that the model only partially reproduced the large water vapour changes seen in MLS data (300–100 hPa) and the model was not able to capture the observed increase in water vapour for the Southern Hemisphere. Despite these issues, the observed water vapour feedback agreed well with previous model estimates and a range of values found from the ensemble of simulations of Pinatubo using the HadCM3 climate model.

We note that our observed value for the water vapour feedback associated with Pinatubo agrees well with the established model estimates associated with long-term climate change. However, to quantify this long-term water vapour feedback in the observations, good observations of water vapour throughout the troposphere are vital over many years. If similarly good observations of long-term sea-ice and cloud changes were available it may also be possible to quantify these feedbacks, employing a similar methodology.

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