Estimates of the Water Vapor Climate Feedback during El Niño–Southern Oscillation

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ABSTRACT

The strength of the water vapor feedback has been estimated by analyzing the changes in tropospheric specific humidity during El Niño-Southern Oscillation (ENSO) cycles. This analysis is done in climate models driven by observed sea surface temperatures [Atmospheric Model Intercomparison Project (AMIP) runs], preindustrial runs of fully coupled climate models, and in two reanalysis products, the 40-yr European Centre for Medium-Range Weather Forecasts Re-Analysis (ERA-40) and the NASA Modern Era Retrospective-Analysis for Research and Applications (MERRA). The water vapor feedback during ENSO-driven climate variations in the AMIP models ranges from 1.9 to 3.7 W m⁻² K⁻¹, in the control runs it ranges from 1.4 to 3.9 W m⁻² K⁻¹, and in the ERA-40 and MERRA it is 3.7 and 4.7 W m⁻² K⁻¹, respectively. Taken as a group, these values are higher than previous estimates of the water vapor feedback in response to century-long global warming. Also examined is the reason for the large spread in the ENSO-driven water vapor feedback among the models and between the models and the reanalyses. The models and the reanalyses show a consistent relationship between the variations in the tropical surface temperature over an ENSO cycle and the radiative response to the associated changes in specific humidity. However, the feedback is defined as the ratio of the radiative response to the change in the global average temperature. Differences in extratropical temperatures will, therefore, lead to different inferred feedbacks, and this is the root cause of spread in feedbacks observed here. This is also the likely reason that the feedback inferred from ENSO is larger than for long-term global warming.

1. Introduction

The water vapor feedback refers to the process whereby an initial warming of the planet, caused, for example, by an increase in atmospheric carbon dioxide, causes an increase in the specific humidity q of the atmosphere. Because water vapor is itself a greenhouse gas, the increase in q causes additional warming.

The water vapor feedback has long been anticipated to exert a powerful warming effect because of the expectation that the atmosphere's relative humidity (RH) would remain roughly constant (Manabe and Wetherald 1967; Randall et al. 2007; Dessler and Sherwood 2009) meaning that q would increase rapidly with surface temperature. Models reproduce this, and it is the single most important reason for the large predicted response of global temperatures to projected increases in greenhouse gases.

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Despite the importance of this feedback, however, there have been just a few efforts to validate its simulation in climate models. Soden et al. (2002) showed that climate models that do not include the water vapor feedback are unable to reproduce observed temperature changes after the eruption of Mount Pinatubo, and Soden et al. (2005) identified a radiative signature of upper-tropospheric moistening over the period from 1982 to 2004 that was captured by climate model simulations. Forster and Collins (2004) used the eruption of Mount Pinatubo to quantitatively estimate the strength of the water vapor feedback and showed that models produced a similar feedback. Minschwaner et al. (2006) and Gettelman and Fu (2008) both used interannual variations to show that observed changes in upper-tropospheric humidity are similar to those seen in climate models.

Climate variations during El Niño–Southern Oscillation (ENSO) events have also been used to analyze the water vapor feedback in climate models. Soden (1997) showed that the observed tropical clear-sky radiative response of the atmosphere during an ENSO cycle compared favorably to that simulated by a model, suggesting that the model's water vapor feedback was reasonable.

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Dessler et al. (2008) calculated the feedback from climate variability from 2003 to 2008, which included ENSO variations, and found a strong, positive feedback. Sun and Held (1996) studied the response of tropospheric water vapor to changes in surface temperature in data and in models and found that the models showed too strong of a response, and Minschwaner et al. (2006) made a conceptually similar comparison (but with different data and more advanced models) and found reasonable qualitative agreement. Sun et al. (2006, 2009) investigated the regional feedback over the tropical Pacific Ocean and found that climate models have a stronger regional water vapor feedback than is indicated by observations. In this paper, we compute the water vapor feedback during ENSO events in a set of climate models and in two reanalysis datasets.

2. Approach

We define $\Delta R[\Delta q(x, y, z)]$ as the change in global average top-of-the-atmosphere (TOA) downward net radiative flux resulting from the change in q at a particular longitude, latitude, and altitude (denoted x, y, and z), with the atmospheric state otherwise held fixed. We write this mathematically as

$$\Delta R[\Delta q(x, y, z)] = \frac{\partial R}{\partial q(x, y, z)} \Delta q(x, y, z), \qquad (1)$$

where $\Delta q(x, y, z)$ is the change in q at a particular longitude, latitude, and altitude and R is the downward global-average TOA net flux. Here, $\partial R/\partial q(x, y, z)$ is the change in R per unit change in q(x, y, z).

In this paper, we use precomputed monthly averaged values of $\partial R/\partial q(x, y, z)$ provided by Soden et al. (2008). Briefly, these are calculated by taking 3-hourly averaged fields of q and other climate variables from a climate model run and perturbing the q field sequentially at each location x, y, and z. Then, $\partial R/\partial q(x, y, z)$ is determined by using a radiative transfer model to calculate how R changes in response to the q(x, y, z) perturbation. Calculations based on 3-hourly fields are then averaged to produce a monthly average value. The radiative computations include clouds simulated by the model.

If Δq is the change in the three-dimensional q field between two climate states, then the total change in Rresulting from Δq is

$$\Delta R_q = \sum_{x,y,z} \Delta R[\Delta q(x, y, z)].$$
(2)

The summation is over all longitudes and latitudes, with appropriate cosine weighting, and altitudes from the surface to 100 hPa. We define the strength of water vapor feedback between these two climate states to be

$$\lambda_q = \frac{\Delta R_q}{\Delta T_G},\tag{3}$$

where ΔR_q is the radiative perturbation defined in Eq. (2) and ΔT_G is the change in globally averaged surface temperature between the two climate states.

In this paper, Δq is the change in q during an ENSO cycle, and we are therefore estimating the strength of the water vapor feedback between the warm El Niño and cool La Niña climate states. We calculate the feedback in Atmospheric Modeling Intercomparison Project (AMIP) model runs, in which an atmospheric model is driven by observed monthly averaged distributions of sea surface temperature and sea ice from 1979 to 2002. We also analyze simulations from fully coupled climate models; here we use preindustrial control runs, in which atmospheric greenhouse gas abundances and other forcings are held constant at preindustrial values. All model runs were obtained from the World Climate Research Programme's (WCRP) Coupled Model Intercomparison Project phase 3 (CMIP3) multimodel dataset (Meehl et al. 2007). Table 1 lists the models used in this analysis.

To test the reality of the feedbacks in these models, we compare the models' feedback to the feedback calculated from the 40-yr European Centre for Medium-Range Weather Forecasts Re-Analysis (ERA-40; Uppala et al. 2005) and from the National Aeronautic and Space Administration (NASA) Modern Era Retrospective-Analysis for Research and Applications (MERRA; Suarez et al. 2008). These reanalysis systems combine a global climate model with observations, and they are therefore not pure observational datasets. However, they have many advantages relative to any single observational dataset and are therefore widely used in place of pure observations in experiments like this one. In this paper, we will consider them as "truth" in our efforts to validate the models. Obviously, how convincing one finds this validation will depend on the confidence one has in the reanalyses.

The ENSO phase in the models and reanalysis systems is determined using the Niño-3 index, defined as the averaged surface temperature anomaly over the area of $5^{\circ}S-5^{\circ}N, 90^{\circ}-150^{\circ}W$. We focus here on the winter months of December–February (DJF), months that are highly affected by ENSO. In the AMIP runs, because the sea surface temperatures in the model are specified from observations, the models' Niño-3 indices are identical to the observed Niño-3 index. For the analysis of these models, we select the strongest ENSO events from the historical record: the El Niño months are December

TABLE 1. The water vapor feedback λ_q and $\Delta R_q / \Delta T_T$ for both AMIP and fully coupled model runs (W m⁻² K⁻¹). The 2σ confidence interval for each quantity is in parentheses. For models with multiple AMIP runs, the calculation includes all of the runs. See Meehl et al. (2007) and references and Internet links therein for details of these models.

| | | AMIP models | | Coupled models | |
|---|------------------------------------------------------------------------------------------------------------|------------------|---------------------------|------------------|---------------------------|
| | Model | λ_q | $\Delta R_q / \Delta T_T$ | λ_q | $\Delta R_q / \Delta T_T$ |
| A | National Center for Atmospheric Research Community Climate System Model, version 3 (CCSM 3.0) | 2.52 (1.98–3.21) | 2.07 (1.77–2.49) | 1.49 (1.21–1.84) | 1.65 (1.38–1.97) |
| В | Goddard Institute for Space Studies (GISS) E20/Russell | 2.60 (2.24–3.06) | 1.91 (1.73–2.16) | 1.38 (1.10–1.72) | 1.58 (1.26–1.96) |
| С | Institute of Numerical Mathematics Coupled Model, version 3.0 (INM-CM 3.0) | 1.92 (1.63–2.22) | 1.83 (1.58–2.37) | 1.92 (1.61–2.34) | 1.68 (1.51–1.86) |
| D | L'Institut Pierre-Simon Laplace Coupled Model, version 4 (IPSL CM4) V1 | 2.38 (2.14–2.68) | 2.23 (2.07–2.40) | 2.89 (2.74–3.07) | 2.25 (2.17–2.34) |
| Е | Model for Interdisciplinary Research on Climate 3.2, high-resolution version [MIROC3.2(hires)] | 2.42 (1.63–3.51) | 1.77 (1.35–2.26) | 2.42 (1.90–3.09) | 1.94 (1.57–2.42) |
| F | Model for Interdisciplinary Research on Climate 3.2, medium-resolution version [MIROC 3.2(medres)] | 2.42 (2.07–2.84) | 1.63 (1.45–1.84) | 2.20 (1.89–2.59) | 1.53 (1.38–1.69) |
| G | ECHAM5/Max Planck Institute Ocean Model (MPI-OM) | 2.95 (2.48–3.53) | 2.39 (2.11–2.69) | 2.75 (2.59–2.91) | 1.72 (1.65–1.80) |
| Η | Meteorological Research Institute Coupled General Circulation Model, version 2.3.2a (MRI CGCM2.3.2a) | 3.75 (3.01–4.62) | 2.21 (1.82–2.81) | 3.94 (3.24–4.60) | 2.11 (1.87–2.39) |
| Ι | Met Office Hadley Centre Global Environmental Model version 1 (UKMO HadGEM1) | 3.24 (2.21–4.65) | 2.21 (1.73–2.83) | 1.78 (1.43–2.21) | 1.88 (1.50–2.35) |
| J | Flexible Global Ocean–Atmosphere–Land System Model gridpoint version 1.0 (FGOALS-1.0g) | 2.25 (2.01–2.50) | 1.79 (1.56–2.05) | 1.54 (1.46–1.63) | 1.58 (1.53–1.63) |
| Κ | ERA-40 | 3.70 (2.42-5.80) | 2.87 (2.27-4.45) | N/A | N/A |
| L | MERRA | 4.71 (2.60-8.11) | 2.49 (1.68–3.33) | N/A | N/A |

1982, January and February 1983, December 1991, January 1992, December 1997, and January and February 1998. The La Niña months are January 1984 and 1985, December 1988, January 1989, December 1998, January and December 1999, and January and February 2000.

For the preindustrial control runs of the fully coupled models, there is no correspondence between ENSO events among the models or between the models and the historical record. Thus, we calculate the Niño-3 index for each individual model and select those DJF months with extreme values. In most cases, we use months with Niño-3 index values above 1.5 or below -1.5, but some models predict few or no months with such large values. For those models, we reduce the threshold until we get a sufficient number of months (30–40) meeting the criterion.

3. Results

Figure 1 shows scatterplots of the TOA radiative anomaly ΔR_q versus global average surface temperature anomaly ΔT_G for the AMIP runs and for ERA-40 and MERRA. Each point represents ΔR_q and ΔT_G for a single El Niño or La Niña month. The Δq used in the calculation of ΔR_q is the difference between q for that month and the average for that month over the entire model time period; ΔT_G is calculated analogously. For models with more than one realization, months from all of the runs are combined in a single scatterplot, which explains why some plots have more points than others.

A linear fit, calculated as the first EOF of the data, is also plotted on each panel. Per Eq. (3), the slope of this linear fit is the strength of the water vapor feedback λ_q . The 2σ confidence intervals, also plotted on each panel, are calculated using a bootstrap method. For each scatterplot, 20 000 new datasets are produced by randomly sampling the original dataset with replacement. A slope is calculated from each random dataset, and the 2σ confidence interval is the range that encompasses 95% of the calculated slopes. In general, the confidence intervals are smaller for those models for which several runs are available, because the number of points going into the fit is largest for these models. For models with more than one run, Table 2 lists the range of λ_q obtained from the individual runs.



FIG. 1. Scatterplots of ΔR_q (W m⁻²) vs ΔT_G (K) for the AMIP climate simulations as well as ERA-40 and MERRA. Also shown is a linear fit to the data, along with the 2σ confidence intervals for the fit. Table 1 lists the models associated with each letter, and Fig. 2a shows the slopes and confidence intervals for each model.

In agreement with previous work, all AMIP models show a positive correlation between ΔR_q and ΔT_G , indicative of a positive water vapor feedback (remember that *R* is defined here as downward positive, so a positive correlation corresponds to a reduction in the radiative flux to space with increasing surface temperature). Figure 2a plots the slope of the fit λ_q for each model, along with the 2σ confidence interval; the numeric values are listed in Table 1. There is a wide range of λ_q among the AMIP models, from 1.9 to 3.7 W m⁻² K⁻¹,

| | Model | λ_q | $\Delta R_q / \Delta T_T$ |
|---|------------------|-------------|---------------------------|
| В | GISS E20/Russell | 2.31-3.13 | 1.79-2.10 |
| D | IPSL CM4 V1 | 2.04-3.10 | 1.99-2.57 |
| F | MIROC3.2(medres) | 2.28-2.53 | 1.54-1.73 |
| G | ECHAM5/MPI-OM | 2.60-3.72 | 2.26-2.48 |
| J | FGOALS-1.0g | 2.07-2.44 | 1.69-1.92 |

with an average of 2.6 W m⁻² K⁻¹ and a standard deviation of 0.53 W m⁻² K⁻¹.

The ERA-40 and MERRA values of λ_q are 3.7 and 4.7 W m⁻² K⁻¹, respectively. The scatter in both the ERA-40 and MERRA plots in Fig. 1 is large, so the confidence interval for λ_q of both are wide—extending from 2.4 to 5.8 W m⁻² K⁻¹ for ERA-40 and from 2.6 to 8.1 W m⁻² K⁻¹ for MERRA. Because the confidence intervals of both overlap with the confidence intervals of

most of the models, we conclude that, within the large uncertainties, the water vapor feedbacks in the models and reanalyses are consistent. However, it is also apparent that the models, taken as a group, predict a weaker water vapor feedback than the reanalyses. Thus, to the extent that the models have deficiencies in their water vapor feedback, the evidence provided by this comparison suggests that the models may be underestimating the feedback.

To better understand the origin of the differences among the models and between the models and the reanalyses, it is first crucial to recognize that the water vapor feedback during ENSO is driven predominantly by changes in tropical mid- and upper-tropospheric q. To see this, Fig. 3a plots

$$\Delta R_q(y) = \sum_{x,z} \Delta R[\Delta q(x, y, z)],$$



FIG. 2. (a) Slope and confidence intervals for the scatterplot of ΔR_q (W m⁻²) vs ΔT_G (K), which is an estimate of the water vapor feedback λ_q , for the AMIP models in Fig. 1, and (b) the slope and confidence intervals for the scatterplot of ΔR_q (W m⁻²) vs tropical temperature anomaly ΔT_T (K). Table 1 lists the models associated with each letter.



FIG. 3. (a) The $\Delta R[\Delta q(y)]$, the change in TOA flux resulting from changes in q at a particular latitude (W m⁻²), and (b) $\Delta R[\Delta q(z)]$, the change in TOA flux resulting from changes in q at a particular altitude [W m⁻² (100 hPa)⁻¹]. In both plots, the models are the gray lines, the ERA40 is the solid black line, and the MERRA is the dashed black line.

the change in global average TOA flux resulting from the change in q at a particular latitude. The plot shows that most of the radiative response to changing q during an ENSO cycle arises from changes in q in the tropics. Dessler et al. (2008) also showed this result (see their Fig. 4). It is due to large changes in q in the tropics over an ENSO cycle combined with high sensitivity of R to changes of q in the tropical upper troposphere (e.g., Soden et al. 2008).

Tropical mid- and upper-tropospheric q, in turn, is determined to a large extent by tropical surface temperatures. Minschwaner and Dessler (2004), for example, were able to successfully explain variations in tropical upper-tropospheric q as a response to variations in tropical surface temperature. Minschwaner et al. (2006) extended this result to show that climate models also manifest this behavior.

Thus, we expect TOA radiative anomalies ΔR_q to scale more closely with the tropical surface temperature

anomaly ΔT_T than, as Eq. (3) suggests, with the global average temperature anomaly ΔT_G . To test this, we have regressed ΔR_q versus ΔT_T . Figure 2b shows the slope of the relation and the 2σ confidence intervals for all models and the two reanalyses.

There is much better agreement among the models and between the models and the reanalysis datasets for the slope of ΔR_q versus ΔT_T than for ΔR_q versus ΔT_G . The $\Delta R_q / \Delta T_T$ from the models ranges from 1.6 to 2.4 W m⁻² K⁻¹, with an average over all models of 2.0 W m⁻² K⁻¹ and a standard deviation of 0.25 W m⁻² K⁻¹. The MERRA value is 2.5 W m⁻² K⁻¹. Considering the confidence intervals of the fits, we see no difference between the MERRA and the models.

The ERA-40 value is 2.9 W m⁻² K⁻¹. This is higher than many models and the MERRA, although its confidence interval overlaps with the MERRA's and those of some of the higher-feedback models. An explanation for this can be found in Fig. 3b, which show that the ERA-40 has larger changes in q in the midtroposphere, around 600 hPa, in the Southern Hemisphere subtropics and midlatitudes that are reproduced by neither MERRA nor the climate models.

Figure 2b shows that the models and the reanalyses are providing a reasonably consistent picture of the relation between global average TOA radiative flux anomalies resulting from changes in q and the corresponding *tropical* surface temperature anomalies. Thus, the primary source of disagreement among the models and between the models and the reanalyses in the magnitude of the water vapor feedback λ_q comes not from differences in how the models handle changes in q, but from differences in the amount of extratropical surface warming over an ENSO cycle.

This is emphasized in Fig. 4, which plots the magnitude of the water vapor feedback λ_q from Fig. 2a against the ratio of $\Delta T_G / \Delta T_T$ for each model. The higher this ratio is, the more extratropical warming per unit tropical warming the model simulates over an ENSO cycle. It is apparent that the models with the highest feedback show the least warming in the extratropics. Less extratropical warming means that ΔT_G is reduced, leading to larger values of λ_q .

We have also performed these analyses for preindustrial control runs of fully coupled versions of the same climate models. Our calculated values of λ_q and the slope of ΔR_q versus ΔT_T for these models are plotted in Fig. 5 and tabulated in Table 1. Overall, there is consistency with the AMIP calculations. The preindustrial control runs show λ_q running from 1.5 to 4 W m⁻² K⁻¹, similar to the range seen in AMIP models, and both show that the slope of ΔR_q versus ΔT_T is around 2 W m⁻² K⁻¹, with much smaller scatter among the models and tighter



FIG. 4. Scatterplot of the feedback strength λ_q for each AMIP model and from the reanalyses against the ratio $\Delta T_G / \Delta T_T$; $\Delta T_G / \Delta T_T$ in this plot is the first EOF of the scatterplot of ΔT_G vs ΔT_T for the ENSO months. Each point is labeled with the letter that indicates which model it is; Table 1 lists the letter associated with each model.

fits (as evidenced by smaller confidence intervals). Thus, our results appear to be robust for various model configurations.

Soden and Held (2006) calculated λ_q for fully coupled models in response to century-long global warming. They found a range of λ_q of 1.6–2.1 W m⁻² K⁻¹, with an average of 1.8 W m⁻² K⁻¹—similar to values obtained by Colman (2003). Taken as a group, the ENSO-derived λ_q calculated here from models is higher than the longterm global warming λ_q . In particular, ERA-40 and MERRA, our closest approximation to truth, are both much higher than the Soden and Held range, with their confidence limits not overlapping.

We thus see evidence that the water vapor feedback in response to ENSO variations is larger than the water vapor feedback in response to long-term global warming. We have not explored in detail the reasons for this difference, but it is certainly possible that at least some of the difference is related to different patterns of surface warming for ENSO and long-term global warming. During ENSO, the warming of the tropics is larger than warming of the extratropics. For long-term global warming, however, we expect extratropical warming to exceed tropical warming. Thus, $\Delta T_G / \Delta T_T$ is larger for centurylong global warming, which tends to produce a smaller feedback.

The feedback calculated here is also slightly larger than that obtained by Dessler et al. (2008), who obtained an average λ_q of 2.0 W m⁻² K⁻¹ from climate fluctuations from 2003 to 2008. The analysis in Dessler et al. confirms, however, that using tropical temperatures instead of global temperatures produces a more consistent estimate of the feedback.

4. Conclusions

We have estimated the strength of the water vapor feedback λ_q in response to climate variations resulting from ENSO in a collection of climate models. The models driven by specified sea surface temperatures (so-called AMIP runs) produce a range of λ_q running from 1.9 to 3.7 W m⁻² K⁻¹, while preindustrial control runs of fully coupled models produce a range running from 1.4 to 3.9 W m⁻² K⁻¹. Two reanalysis systems—ERA-40 and MERRA—produce values of 3.7 and 4.7 W m⁻² K⁻¹, respectively. The feedback found in fully coupled preindustrial control runs is similar to that found in the AMIP runs.

The comparison between the water vapor feedback in the models and reanalysis, which is the main point of the paper, produced intriguing results. Taken as a group, the models tend to *underpredict* the ENSO-driven water vapor feedback found in the reanalyses. However, the confidence intervals on the reanalyses in particular are large, so we cannot exclude agreement.

The models and MERRA show good agreement on the response of global average TOA radiative flux anomalies ΔR_q to anomalies of the tropical surface temperature ΔT_T . This supports the idea that the water vapor feedback during ENSO is primarily caused by changes in tropical q, which in turn is primarily regulated by tropical surface temperature. Our analysis suggests that the models simulate this process well. ERA-40 shows a larger response in ΔR_q owing to a larger change in midtropospheric q in the Southern Hemisphere subtropics and midlatitudes.

Thus, disagreements in the ENSO-driven water vapor feedback among the models and reanalyses are primarily



FIG. 5. (a) Slope and confidence intervals for the scatterplot of ΔR_q (W m⁻²) vs ΔT_G (K), which is an estimate of the water vapor feedback λ_q , for the preindustrial control runs, and (b) the slope and confidence intervals for the scatterplot of ΔR_q (W m⁻²) vs tropical temperature anomaly ΔT_T (K). Table 1 lists the models associated with each letter.

due to differences in how much extratropical surface temperature varies over an ENSO cycle. Models with large extratropical warming per unit tropical warming will have lower values of λ_q than models with small extratropical warming.

Thus, the spread of the feedbacks among the models and between the models and the reanalyses is a consequence of the assumption implicit in the definition of the water vapor feedback that it is proportional to global average temperature. This also is the likely explanation of why the calculated feedback during ENSO is larger than it is for long-term global warming.

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