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Are climate-related changes to the character of global-mean precipitation predictable?

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Abstract

The physical basis for the change in global-mean precipitation projected to occur with the warming associated with increased greenhouse gases is discussed. The expected increases to column water vapor W control the rate of increase of global precipitation accumulation through its affect on the planet's energy balance. The key role played by changes to downward longwave radiation controlled by this changing water vapor is emphasized. The basic properties of molecular absorption by water vapor dictate that the fractional rate of increase of global-mean precipitation must be significantly less that the fractional rate of increase in water vapor and it is further argued that this reduced rate of precipitation increase implies that the timescale for water re-cycling is increased in the global mean. This further implies less frequent precipitation over a fixed period of time, and the intensity of these less frequent precipitating events must subsequently increase in the mean to realize the increased global accumulation. These changes to the character of global-mean precipitation, predictable consequences of equally predictable changes to W, apply only to the global-mean state and not to the regional or local scale changes in precipitation.

Keywords: global precipitation, climate change

1. Introduction

The Clausius–Clapeyron relation states that the saturation partial pressure of water vapor increases roughly as an exponential function of temperature. The observed water vapor content of the atmosphere, measured as the vertical integral of specific humidity (hereafter the column water vapor W), also follows this approximate exponential relation with respect to surface temperature (Stephens 1990). The Clausius–Clapeyron relation also predicts the partial pressure increases at a rate of about 6% K^{-1} at 300 K and 15% K^{-1} at 200 K. The increase in W in model simulations of climate warming also occurs at the rate of approximately 7% K^{-1} (e.g. Held and Soden 2006, among others) and thus seems to come under the influence of this simple thermodynamic control. Satellite observations appear to support these model-predicted rates of increase of W (Trenberth *et al* 2005, Santer *et al* 2007) as do observed increases in surface humidity over the past few decades (Willet *et al* 2007). Other observational studies, however, note that changes in water vapor are more complicated than is suggested by this simple picture with increases that occur low in the atmosphere being coupled to apparent changes in stratospheric vapor (Solomon *et al* 2010) as well as to changes in mid-to-upper tropospheric water vapor (e.g. Paltridge *et al* 2009 and Soden *et al* 2005).

Model-predicted changes to global precipitation have the following general attributes: (i) the global-mean precipitation increases at a rate (∼2% K^{-1}) that is significantly less than the predicted increase of water vapor (7% K^{-1}) (ii) precipitation does not increase everywhere uniformly, and the distributions of change does not simply follow the patterns of water vapor change, with increase occurring primarily in raining regions and decreases occurring in existing arid regions (e.g. Meehl *et al* 2005, Neelin *et al* 2006), and (iii) the frequency of

Table 1. The global, annual mean surface and TOA energy fluxes as summarized in the study of Trenberth *et al* (2009).

	$R_{sw,net}$ (W m^{-2})	$R_{lw,net}$ (W m^{-2})	LE (W m^{-2})	S (W m^{-2})	ε (W m^{-2})
Surface	161	−63 (=−396 + 333)	80	17	0.9
Top-of-atmosphere (TOA)	239 (=341 − 102)	239			

precipitating events decreases overall although the heavier rain events appear to become more frequent (e.g. Pall *et al* 2007). Observational evidence (e.g. Trenberth *et al* 2007) supporting these findings are mixed. Most of the observations that seem to support these model projections are not global, being restricted to over land (e.g. Zhang *et al* 2007), or limited to the tropics (Allan and Soden 2007). Sun *et al* (2006), for example, suggest most models reproduce the spatial patters of precipitation frequency and intensity over land although the models tend to rain too frequently at reduced intensity as noted in other studies (e.g. Dai and Trenberth 2004). The only truly near-global (land + ocean) data source of precipitation is that of GPCP but these data are not truly homogeneous and interpretation of any trend in this relatively short time series of data requires caution. Gu *et al* (2007) note, for instance, that 'the global linear change of precipitation is near zero' (less than 1% K^{-1} based on their published trends) yet Wentz *et al* (2007) using their own satellite microwave based precipitation product over oceans combined with the over land GPCP precipitation arrive at an entirely conflicting result with precipitation changes approaching 6% K^{-1} over the past two decades. That such differences exist merely highlights the inconsistencies in these global data sources themselves and serves as reminder that declaring trends in relatively short time records of data appears premature.

The physical basis for certain expectations of change in global-mean precipitation associated with global warming is discussed and it is argued these global changes are predictable to the extent that increases in water vapor W are predictable. Three relevant aspects of the character of precipitation are the focus of this argument, and are introduced with the following notional expression:

$$\begin{aligned}&\text{Accumulation over some fixed time}\\&\quad = \Sigma(\text{frequency of precipitation})\\&\qquad \times (\text{intensity of precipitation when it occurs}).\end{aligned} \tag{1}$$

It will be explained (i) how the expected changes to global-mean W in fact control the rate of change in global-mean precipitation accumulation (ii) why this increase must be proportionally less than the increase in W with the implication for an increased water re-cycling time and less frequent precipitation, and thus (iii) why the intensity of precipitating events in the mean are expected to increase.

2. The Earth's energy balance

That changes in global precipitation are controlled by changes in the Earth's global-mean energy balance has been appreciated for some time (Allen and Ingram 2002, Stephens 1999, Stephens and Ellis 2008, among several others). To understand how this energy balance changes in a climate system forced by increasing CO_2, we first consider the energy balance of the current climate as reviewed recently by Trenberth *et al* (2009). Their estimates of the global fluxes are a mix of observations and model data and averaged over the 2002–2007 period are given in table 1.

The surface energy balance is

$$R_{sw,net} + R_{lw,net} = \text{LE} + S + \varepsilon \tag{2}$$

where $R_{sw,net}$ and $R_{lw,net}$ are the net shortwave and longwave fluxes into the surface and $\varepsilon = 0.9$ W m^{-2} is supposed to represent the heat stored in the ocean during for the 2002–2007 period. While the top-of-atmosphere (TOA) fluxes obtained from satellite observations have well characterized errors (Wielicki *et al* 1996), the errors attached to surface fluxes are less well known but appear to be large for some components. The downward longwave radiative flux (DLR) in particular (333 W m^{-2}), is significantly lower by 10–20 W m^{-2} than other flux estimates available from different observational and model data sources that have been assembled under the GEWEX radiative flux assessment activity (Stephens *et al* 2010). This suggests significant biases also exist in the other surface fluxes. This difference is of some relevance to the topic of this paper as it is argued that changes to the DLR exert the most influential control on the change in global-mean precipitation. Despite the significant uncertainty in DLR, it is reasonable to assume that the uncertainties in the likely change to DLR due with climate warming are significantly smaller than the uncertainty on DLR itself.

3. Forced changes to the Earth's energy balance

Changes to the net TOA radiative fluxes associated with climate warming forced by increasing CO_2 are too small to be detectable by present day observing systems. Significant compensating changes to the radiative fluxes, however, occur within the atmosphere and at the surface in such a way that changes to the net surface radiative fluxes merely mirror the equivalent changes to atmosphere radiative heating. Climate model depictions of these changes are revealed in figures 1 and 3. Figure 1(a) is a plot of the change in TOA outgoing longwave fluxes (OLR) taken from the outputs of a selection of coupled climate model simulations. The model data are from the archives of the World Climate Research Programme's (WCRP's) coupled model inter-comparison project phase 3 (CMIP3, Meehl *et al* 2007) multi-model data set (www-pcmdi.llnl.gov/) for the 1% per year increase in CO_2 scenario experiments as summarized in the Intergovernmental Panel on Climate Change Fourth Assessment Report (IPCC AR4) (www.ipcc.ch). The resultant change in the DLR (ΔDLR) is given in figure 1(b). The fundamental explanation for why the

Table 2. Calculated changes in DLF and net surface longwave flux that arise from atmospheric temperature and water vapor perturbations as described in the text.

Atmospheric model	T_s (K)	Column water vapor (kg m^{-2})	ΔDLR$_{\Delta T}$ (W m^{-2})	ΔDLR$_{\Delta T+\Delta q}$ (W m^{-2})	ΔF_{SFC} (W m^{-2})	$\Delta R_{net,LW}$ (W m^{-2})
Tropical	300	41.2	4.7	10.1	6.1	−4.0
Mid-lat summer	294	29.3	4.2	9.0	5.8	−3.2
Mid-lat winter	272.2	8.6	3.1	4.6	4.6	0.0
Sub-arctic summer	287.0	20.8	3.7	7.3	5.4	−1.9
Sub-arctic winter	257.1	4.2	2.5	3.5	3.9	0.4

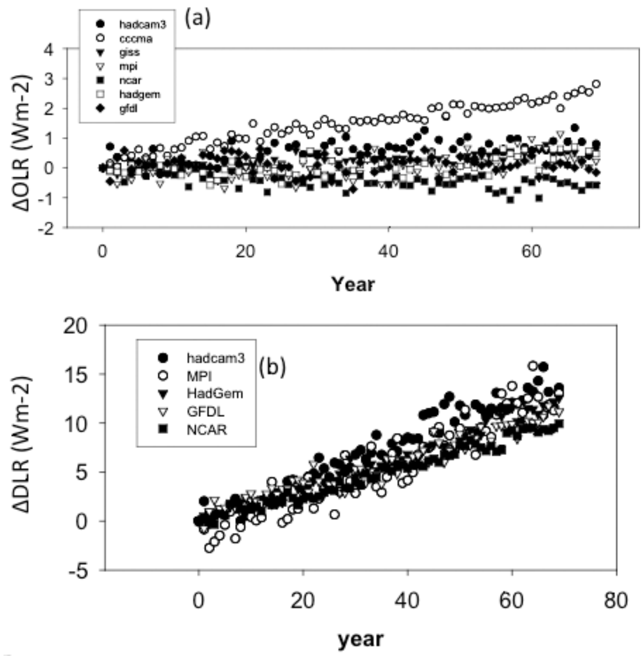

Figure 1. (a) Change in OLR as a function of time from present CO_2 levels from CMIP-3 1% per year increase in CO_2 scenario experiments. (b) As in (a) but for the change in DLF.

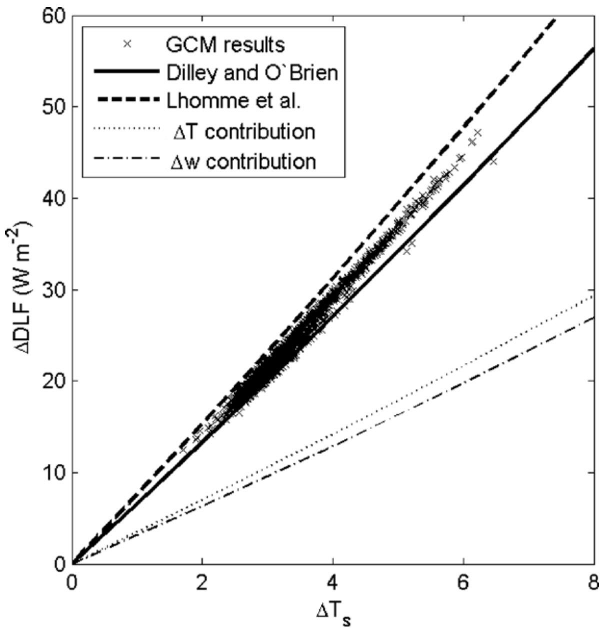

Figure 2. Changes in all-sky DLR (symbols) as a function of global-mean surface temperature change obtained from the CO_2 climate change experiments of http://climateprediction.net. The heavy solid and dashed lines are the projected changes in clear-sky DLR according to two different parameterizations of the DLR as heavy solid and dashed lines. The individual contributions to this projected change predicted by the Dilley and O'Brien model due to the warming and moistening of the atmosphere are also shown.

OLR changes are small while the surface DLR changes are amplified is described in Stephens (1999).

The contribution to ΔDLR by increased CO_2 is small (about 1 W m^{-2} with a doubling of CO_2, Ramanathan 1981, Stephens and Ellis 2008) in comparison to the changes shown in figure 1(b) and the factors that determine the magnitude of the changes shown can be understood by reference to table 2. Listed on this table are changes in DLR calculated using a broad-band radiative transfer model (Stephens *et al* 2001) applied to the five different McClatchey *et al* (1972) model atmospheres characterized by the surface temperatures and column water vapor given for reference. Listed are values ΔDLR due to a 1 K change in atmospheric temperature (at all levels, ΔDLR$_{\Delta T}$) and the combined change in DLR for a 1 K warming and an assumed 7% increase in water vapor throughout the column (ΔDLR$_{\Delta T+\Delta q}$). The difference in values between these two columns indicates the effect of water vapor on ΔDLR and we can conclude that the change in DLR shown in figure 1(b) arises from almost equal changes in water vapor and temperature (see also discussion of figure 2). The table also shows the change in (upward) surface emission of longwave flux due to a 1 K surface warming and the final column is the change in net surface longwave flux (differences of columns 7 and 6) that is primarily controlled by changes in W (also figure 3(a) and discussion) since the temperature based changes in upward longwave flux approximately cancels the temperature component of the ΔDLR. The net shortwave flux changes at the surface are also, to first order, a function of the change in W (figure 3).

The above discussion of table 2 provides a basis for the interpretation of the surface radiative flux changes summarized in figures 2 and 3. Figure 2 presents changes in all-sky DLR as a function of changes in global-mean surface temperature. Also shown are the predicted changes to clear-sky DLR predicted by two different algorithms (Lhomme *et al* 2007, Dilley and O'Brien 1998) using the given changes of temperature and water vapor taken from the model simulations. Also shown are the separate contributions to this clear-sky change due to the temperature and water vapor influences on DLR. Figure 3(a) shows the companion changes in net SW, LW

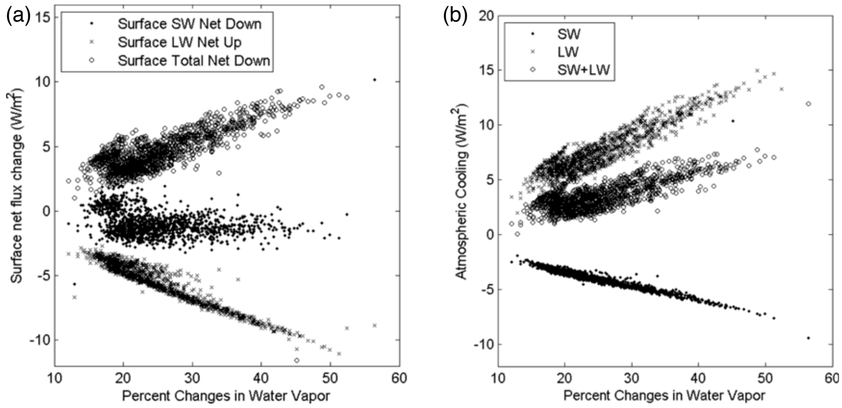

Figure 3. (a) Changes in net SW, LW and total (SW + LW) net surface fluxes as a function of the per cent change in water vapor derived from the same CO_2 climate change experiments of http://climateprediction.net presented in figure 2. (b) As in (a) but for change in atmospheric radiative heating.

and total (SW + LW) net surface fluxes as a function of the per cent change in water vapor. Figure 3(b) is the equivalent figure for the changes in the components of the atmospheric radiation balance and these components are primarily a mirror image of the corresponding surface fluxes as noted above. All data shown in figures 2 and 3 are derived from the transient CO_2 climate change experiments of http://climateprediction.net based on a version of the Hadley center coupled ocean–atmospheric model (HadCM3). The changes are from similar transient experiments as in figure 1 and include outputs from 1380 different models with perturbed parameterizations of physical processes.

The key points drawn from the results displayed in these figures are: (i) model changes in all-sky DLR (figure 2) are primarily governed by changes in clear-sky DLR as indicated by the degree to which the parameterizations match the model data. We also infer that the changes of global-mean DLR are almost equally split between the temperature and water vapor effects on the DLR; (ii) the surface radiative flux changes of figure 3(a) mirror the atmospheric heating changes of figure 3(b) as expected for a system close to equilibrium; (iii) the most dominant change in $\Delta R_{\mathrm{atm,net}}$ (and thus $\Delta R_{\mathrm{sfc,net}}$) occurs as a result of changes in DLR as shown in figure 1(b) and supported by the results of figure 2. (iv) Both the net longwave and shortwave flux changes at the surface are also, to first order, a function of the change in W as revealed in figure 3(b) and are controlled by changes in water vapor (also table 2). Thus $\Delta R_{\mathrm{atm,net}}$ is a strong function of the changes in water vapor.

The implications of these results are now explored by considering the change in radiative fluxes expressed in terms of a change in the net clear-sky fluxes and a change in the cloudy-sky fluxes

$$\Delta R_{\mathrm{atm,net}} = \Delta R_{\mathrm{atm,clr}} - \Delta C_{\mathrm{net}}. \tag{3}$$

The clear-sky portion of this net flux contains a term that varies as a simple power law of W of the form

$$R_{\mathrm{atm,clr}} \sim a W^b \tag{4}$$

where for simplicity and for clarity only we have ignored (obvious) contributions to $R_{\mathrm{atm,clr}}$ by other greenhouse gases and do so without loss of relevance to this discussion (Stephens and Ellis 2008 provide a more complete version of (4) with these other factors represented). Equation (4) simply implies that

$$\frac{\Delta R_{\mathrm{atm,net,clr}}}{R_{\mathrm{atm,net,clr}}} \approx b\frac{\Delta W}{W} \tag{5}$$

and thus a fractional change in atmospheric radiative heating equates directly to a fractional change in column water vapor W scaled by the exponent b.

The simple power-law relation (4) requires further comment. A fundamental property of absorbing gases is the 'curve-of-growth' (e.g. Chamberlain and Hunten 1987) in which the absorption (and emission) increases at a decreasing rate as the absorbing gas path increases. This logarithmic dependence of absorption on absorbing gas amount is a fundamental property of spectral line absorption properties that dictates $b = 0.5$ (the so-called square-root law for single absorbing lines). When considering both solar and infrared effects and integrating over broad spectral regions of overlapping lines, $b < 0.5$. Thus we can assert that as a consequence of the basic properties of molecular absorption, $b < 1$.

4. The global relation between changes of energy and precipitation

At the surface we write the changes to the energy balance (2) as

$$\Delta R_{\mathrm{sw,net}} + \Delta R_{\mathrm{lw,net}} = \Delta \mathrm{LE} + \Delta S, \tag{6}$$

where changes to fluxes of latent (ΔLE) and sensible (ΔS) heat balance changes that occur to surface radiation fluxes ($\Delta R_{\mathrm{sw,net}} + \Delta R_{\mathrm{lw,net}}$) and changes to heat storage are ignored. As noted for an equilibrium system, the latter are equivalent to changes in the radiation budget of the atmosphere ($\Delta R_{\mathrm{atm,net}}$) given balance at the TOA ($\Delta R_{\mathrm{TOA,net}} = 0$). Therefore changes in $\Delta R_{\mathrm{atm,net}}$ forced by changes to CO_2 but controlled by water vapor changes are balanced by changes in surface sensible and latent heating on the global-mean scale.

The relation between changes to the energy balance as presented above and changes to the accumulated precipitation simply follow by replacing ΔLE with $L\Delta P$ where ΔP is the change in global accumulated precipitation. It then follows from a combination of (6) and (3) that

$$\Delta R_{\rm atm,net,clr} - \Delta C_{\rm net} = L\Delta P + \Delta S$$

and by ignoring both $\Delta C_{\rm net}$ and ΔS for the moment, it follows from (3) and (4) that

$$\frac{\Delta R_{\rm atm,net,clr}}{R_{\rm atm,net,clr}} = \frac{\Delta P}{P} \approx b\frac{\Delta W}{W} \tag{7}$$

which further implies that

$$\frac{\Delta P}{P} < \frac{\Delta W}{W} \tag{8}$$

since $b < 1$. That is, under the assumption that $\Delta C_{\rm net}$ and ΔS are second order influences on the energy balance, the fractional rate of increase of the globally accumulated precipitation is significantly less than the equivalent fractional rate of increase of global water vapor. Again this inequality is dictated by the basic properties of molecular absorption and we refer to this controlling influence on ΔP is hereafter referred to as 'the curve of growth' effect.

The two remaining flux terms, $\Delta C_{\rm net}$ and ΔS, have been ignored in this simple analysis but they cannot *a priori* be assumed negligible although they can be expected to be smaller than the curve of growth term that dominates the changes in surface radiation balance (figure 3(a)). Figure 4 is a simple summary of the magnitudes of change of the different factors that contribute to the rate of total change of accumulated precipitation derived from the models analyzed in Stephens and Ellis (2008). The fluxes of sensible heat decrease in all models and this decrease, with all other factors remaining the same, requires further increases in precipitation to offset its reduced effects on the energy balance. This reduction of sensible heating is significantly smaller than the curve of growth term and both its magnitude and sign of influence on precipitation are a consequence of predictable decreases in the difference between surface temperature and the air temperature immediately above the surface (e.g. Richter and Xie 2008, Lu and Cai 2009). Unlike the sensible heating term, the cloud heating term has *a priori* neither an obvious predictable sign nor predictable magnitude. For the climate model simulations analyzed in Stephens and Ellis (2008), changes to clouds in all models induced a net heating of the atmosphere which involve high cloud changes that in turn reduce the enhanced radiative loss associated with increased water vapor. This effect in turn reduces the global precipitation change required for energy balance (hence the negative contribution in figure 4). For the models considered, both $\Delta C_{\rm net}$ and ΔS are of approximately equal magnitude but opposite sign, and cancel giving the net change in this case determined almost entirely by the curve-of-growth effect. There is no reason to expect that such cancelation is a basic property of the real climate system.

5. Discussion

One of the important consequences of the inequality (7) is that the timescale of cycling of water through the atmosphere must

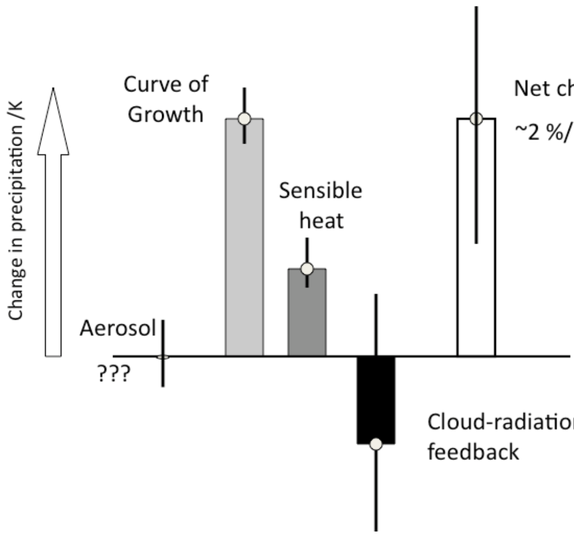

Figure 4. The contributions to rate of change of global precipitation (% K^{-1}) diagnosed from multi-model climate simulations (adapted from Stephens and Ellis 2008). The error bars are qualitative depictions of relative uncertainties of each term.

increase with global warming (e.g. Bosilovich *et al* 2005). For the multi-model data analyzed by Stephens and Ellis (2008), this re-cycling timescale increases by almost a day (an increase of $\sim$10%) at the approximate time of doubling CO_2. One of the ways this slowing of the water re-cycling can occur is through a reduction in the mean frequency of precipitation. According to (1), an increase in global accumulation of precipitation that occurs with an associated decrease in mean precipitation frequency implies that the intensity of precipitating events must, in the mean, increase. These inferred changes to the mean precipitation accumulation, frequency and intensity, we argue, are a direct consequence of an increasing global-mean W.

The control on global-mean precipitation that occurs as a result of the way changes in W affect the planetary energy balance described above neither offers insight on precisely what mechanisms produce these changes nor guides understanding of possible regional shifts in precipitation. Several studies point out that regional changes in precipitation are heterogeneous and do not conform the global-mean changes described above. The 'rich get richer and the poor get poorer' as noted by Neelin *et al* (2006) and discussed further by Trenberth (2009). Pall *et al* (2007) note how the regions of heaviest precipitation appears to increase at a rate similar to the increase in water vapor suggesting that on this scale the increase in moisture convergence into such weather systems is an essential mechanism in the local climate responses (e.g. Trenberth *et al* 2003).

Richter and Xie (2008) and Lu and Cai (2009) provide a more mechanistic view of the changes in surface evaporation that occurs in the AR4 multi-model ensemble simulations. Richter and Xie (2008) find that near surface relative humidity increases of 1.0%, near surface stability increases, as measured by air–sea temperature difference, of 0.2 K (i.e. a reduction in near surface air temperature that is also responsible for the sensible heat flux decrease), and near surface wind speed decreases by 0.02 m s^{-1} combine to produce increases in surface evaporation over ice-free oceans evaporation of only 2% K^{-1}. Lu and Cai (2009) similarly propose that the increase in atmospheric boundary layer stability is the main mechanism responsible for the simultaneous reduction of the surface sensible flux and the smaller than expected increase in surface latent heat flux. There is also a dynamical aspect to the mechanism by which the energy balance controls are exerted. Held and Soden (2006) suggest that reduction in convective mass fluxes, consistent with the atmospheric stability changes noted in Lu and Cai (2009), lead to smaller increases in the rate in precipitation in a warmer climate. This reduction in convective mass flux is associated with the weakening of tropical Walker circulation noted in the climate models (Vecchi *et al* 2006).

It is tempting to look to other effects on the energy balance that might alter the inequality between the rates of increase of water vapor and precipitation. One obvious possibility is the effect of changing aerosol on the net solar radiation into the surface ($\Delta R_{\mathrm{sw,net}}$) as proposed, for example, by Previdi and Liepert (2008). The eruption of Mt Pinatubo in June 1991, for instance, increased the reflected solar radiation in the tropical latitudes. This increased reflected radiation perturbed the planet's radiative equilibrium and was followed by a decrease in precipitation over land (Trenberth and Dai 2007). It is a mistake, however, to assume that the response of a system in equilibrium is the same as that of a non-equilibrium system as in the Mt Pinatubo example where changes to surface $\Delta R_{\mathrm{sw,net}}$ that occurred from the increase in scattering aerosol are balanced primarily by changes to $L\Delta P$. These changes to surface solar fluxes when they are due to increased scattering by aerosol are also typically mirrored by changes in the net solar flux at the TOA as observed after the Mt Pinatubo eruption. In fact this connection between surface solar changes and TOA changes is the basis for the retrieval of surface solar fluxes from TOA measured fluxes (e.g. Li *et al* 1993). Unlike the non-equilibrium Pinatubo example, an equilibrium response requires energy that changes to the TOA net solar drive changes in the OLR that further requires that the longwave fluxes at the surface (and within the atmosphere) must also adjust. It is this adjustment that in turn primarily alters $L\Delta P$ for such an equilibrium system. Under these circumstances, it is to be expected that the effects of scattering aerosol on the rate of change of global-mean precipitation with temperature are likely to small compared to the effects of the other terms shown in figure 4. Absorbing aerosol by contrast can change the global atmospheric radiation balance, reducing the atmospheric cooling and thus reducing the global precipitation as has been demonstrated in climate model simulations (e.g. Lohmann 2007). Although absorbing aerosols can provide significant regional absorption of solar radiation, the magnitude of the influence of absorbing aerosol on the global energy balance is not known. It is reasonable to suggest that the effects of these absorbing aerosols on global precipitation are likely to be smaller than the other factors summarized in figure 4.

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