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How closely do changes in surface and column water vapor follow Clausius–Clapeyron scaling in climate change simulations?

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Abstract

The factors governing the rate of change in the amount of atmospheric water vapor are analyzed in simulations of climate change. The global-mean amount of water vapor is estimated to increase at a differential rate of 7.3% K^{-1} with respect to global-mean surface air temperature in the multi-model mean. Larger rates of change result if the fractional change is evaluated over a finite change in temperature (e.g., 8.2% K^{-1} for a 3 K warming), and rates of change of zonal-mean column water vapor range from 6 to 12% K^{-1} depending on latitude.

Clausius–Clapeyron scaling is directly evaluated using an invariant distribution of monthly-mean relative humidity, giving a rate of 7.4% K^{-1} for global-mean water vapor. There are deviations from Clausius–Clapeyron scaling of zonal-mean column water vapor in the tropics and mid-latitudes, but they largely cancel in the global mean. A purely thermodynamic scaling based on a saturated troposphere gives a higher global rate of 7.9% K^{-1}.

Surface specific humidity increases at a rate of 5.7% K^{-1}, considerably lower than the rate for global-mean water vapor. Surface specific humidity closely follows Clausius–Clapeyron scaling over ocean. But there are widespread decreases in surface relative humidity over land (by more than 1% K^{-1} in many regions), and it is argued that decreases of this magnitude could result from the land/ocean contrast in surface warming.

Keywords: water vapor, climate change, global warming, hydrological cycle, precipitation

1. Introduction

Increases in the amount of atmospheric water vapor under global warming are of climatic importance because of water vapor's role in energy transport by latent heat fluxes, patterns of precipitation and evaporation, radiative transfer, and freshwater exchange with the ocean (Peixoto and Oort 1992). The increase in the amount of water vapor for a given temperature change is strongly constrained by the Clausius–Clapeyron relation. This gives a fractional rate of change of saturation vapor pressure that varies substantially over the range of typical tropospheric temperatures: from ∼6% K^{-1} at 300 K to ∼15% K^{-1} at 200 K.[1]

Rates of change of column water vapor are frequently cited as a baseline quantification of changes in the amount of water in the atmosphere, especially because satellite observational estimates are available for this quantity[2]. Held and Soden (2006) found a rate of increase in the amount of global-mean water vapor with respect to global-mean surface

[1] The moist-thermodynamic formulation used is described in section 3.
[2] Changes in upper-tropospheric water vapor have been extensively studied because of their importance for radiative transfer, but are not addressed here.

air temperature of 7.5% K^{-1} based on a range of simulations with different climate models. Several authors have argued that such a rate of change of global-mean water vapor is consistent with the rate of change of saturation vapor pressure at a typical lower-tropospheric temperature, given that water vapor is mostly concentrated near the surface and relative humidity does not change greatly in climate change simulations. For example, Trenberth *et al* (2003) found a fractional rate of increase in saturation specific humidity of about $\sim$7% K^{-1} to be representative based on consideration of the Clausius–Clapeyron relation and global-mean temperatures at pressures of 700 and 850 hPa. Similarly, column water vapor has been found to vary with surface temperature at a rate of $\sim$9% K^{-1} in observations over tropical oceans, and this has been related to Clausius–Clapeyron scaling by assuming a constant rate of change with respect to a lower-tropospheric temperature, and by relating the lower-tropospheric and surface temperature variations using a constant factor related to the moist-adiabatic lapse rate (Wentz and Schabel 2000). The common approach of picking a representative lower-tropospheric level at which to evaluate the Clausius–Clapeyron rate of change of column water vapor is a reasonable first approximation, but it is not sufficiently exact to allow the quantification of the various contributions to the changes in column water vapor. It also does not take direct account of the greater degree of tropical warming at higher levels and greater fractional rate of change of saturation vapor pressure with respect to temperature at higher levels.

The first purpose of this note is to directly calculate the rate of change of column water vapor under climate change that would result from Clausius–Clapeyron scaling. This is important given the frequency with which Clausius–Clapeyron scaling is identified with 6.5 or 7% K^{-1} in the literature, with values above 7% K^{-1} sometimes referred to as exceeding Clausius–Clapeyron scaling. We will primarily interpret Clausius–Clapeyron scaling as corresponding to the fractional rate of change of water vapor that results from an invariant monthly-mean distribution of relative humidity. We will also discuss the rate of change of water vapor that would be experienced by a saturated troposphere, since this gives a purely thermodynamic scaling. Our calculations are based on climate model simulations, which can provide the necessary distributions of mean temperature, temperature changes, and mean relative humidity (cf Mears *et al* 2007).

The second purpose of this note is to more generally determine the factors governing the calculated rates of change of column water vapor and surface specific humidity in simulations of global warming scenarios. The rate of change of tropospheric column water vapor may be affected by changes in mean relative humidity, which although smaller than changes in mean specific humidity, are nonetheless expected and follow systematic geographical patterns in climate model simulations (Mitchell and Ingram 1992, Lorenz and DeWeaver 2007, Sherwood *et al* 2010). Surface humidity is climatically important for a number of reasons, including its role in the surface energy budget and in moist convection. Surface specific humidity and column integrated water vapor may behave differently because of the variation with height in mean temperatures and temperature changes, and because surface relative humidity over ocean may be more tightly constrained by energetics than relative humidity in the free troposphere (Held and Soden 2000, Schneider *et al* 2010), and because surface relative humidity over land can be expected to be directly influenced by moisture availability limitations on evaporation rates. Lastly, if the fractional rate of change in the amount of water vapor is calculated over a finite temperature change, then it will be greater the larger the temperature change, since the amount of water vapor is expected to increases quasi-exponentially with increasing temperature.

We begin by quantifying the effect of finite temperature changes on calculated rates of change in the amount of water vapor (section 2). Turning to model simulations (section 3), we compare the rates of change of column water vapor with those given by Clausius–Clapeyron scaling at different latitudes (section 4). We also compare the fractional rates of change of column water vapor with those of surface specific humidity, and discuss a possible cause for decreases in surface relative humidity over land under global warming (section 5). Our conclusions include a brief discussion of the implications of our results for precipitation rates (section 6).

2. Effect of finite temperature changes

Fractional rates of change of water vapor in climate change simulations are often calculated as

$$r_\Delta = \frac{c_2 - c_1}{c_1 \Delta T}, \tag{1}$$

where c_i is the column water vapor (or any other measure of the amount of water vapor), $\Delta T = T_2 - T_1$ is the change in temperature, and $i = 1, 2$ correspond to two different climate states (e.g., Boer 1993, Held and Soden 2006). We can also define a differential fractional rate of change as $r = \mathrm{d} \log c / \mathrm{d}T$, where log is the natural logarithm (cf Lorenz and DeWeaver 2007). Because of the roughly exponential dependence of water vapor on temperature, the finite difference estimate will generally be an over-estimate of the differential fractional rate of change, that is, $r_\Delta \geqslant r$ (O'Gorman and Schneider 2008). Making the simplifying assumption that water vapor amounts depend exponentially on temperature for small enough temperature changes, we can easily convert between the finite difference (r_Δ) and differential (r) rates of change using

$$r = \frac{\log(1 + r_\Delta \Delta T)}{\Delta T}. \tag{2}$$

The assumption of an exponential dependence on temperature is not quite correct—for example, the Clausius–Clapeyron dependence on temperature is not exactly exponential—but it should be adequate to capture the leading-order correction for finite temperature changes.

Equation (2) is useful in that it allows the comparison of rates of change in different climate change scenarios[3]. We will report our results as differential rates of change (r), calculated

[3] Lenderink and van Meijgaard (2008) also accounted for the effect of finite temperature changes, but did so by renormalizing to a temperature change of 1 K.

by applying equation (2) to the finite difference rates of change. The magnitude of the correction involved is illustrated using the example of the rate of change of global water vapor discussed in the next section: a differential rate of change of $r = 7.3\%\ \mathrm{K}^{-1}$ corresponds to $r_\Delta = 7.6\%\ \mathrm{K}^{-1}$ for $\Delta T = 1$ K, $r_\Delta = 8.2\%\ \mathrm{K}^{-1}$ for $\Delta T = 3$ K, and $r_\Delta = 9.2\%\ \mathrm{K}^{-1}$ for $\Delta T = 6$ K.

3. Models simulations and analysis

We analyze simulations from the World Climate Research Programme's (WCRP's) Coupled Model Intercomparison Project phase 3 (CMIP3). The identifiers of the models used are: BCCR-BCM2.0, CNRM-CM3, CSIRO-MK3.5, GFDL-CM2.0, GFDL-CM2.1, IAP FGOALS-G1.0, IPSL-CM4, MIROC3.2 (hires), MIROC3.2 (medres), MPI-ECHAM5, MRI-CGCM2.3.2, and CCSM3. Results are presented based on differences between time averages over the final 20 years of the 20th century (1980–1999) in the 20C3M emissions scenario and of the 21st century (2080–2099) in the A1B emissions scenario.

Values of saturation specific humidity are not reported in the model archive, and so we calculate the saturation vapor pressure according to a modified Tetens formula (Simmons *et al* 1999), as the saturation vapor pressure over ice for temperatures below $-23\,^\circ$C, the saturation vapor pressure over liquid water above $0\,^\circ$C, and a quadratic interpolation between the two at intermediate temperatures. All calculations are based on reported monthly-mean temperatures and specific humidities.

Changes in column water vapor corresponding to Clausius–Clapeyron scaling are evaluated as the change in vertically integrated specific humidity assuming that the seasonally varying distribution of mean relative humidity remains invariant (cf Soden *et al* 2005). In other words, the value for the 20th century is simply the mass-weighted vertical integral of specific humidity; but the value for the 21st century is the mass-weighted vertical integral of a specific humidity computed from the seasonally varying mean relative humidity over the 20th century time period and the 21st century saturation vapor pressure. We also calculate a purely thermodynamic scaling for column water vapor as the change in the vertically integrated saturation specific humidity over the troposphere. This depends only on temperature and pressure and will be referred to as the change in saturation column water vapor.

In all cases, vertical integrations are taken over the troposphere to avoid problems with saturation specific humidities at low temperatures. Specific humidities and saturation specific humidities are computed pointwise (for each month, pressure level, latitude and longitude), and then vertically integrated from the surface to the zonal-mean tropopause. The tropopause is defined as a level with lapse rate of monthly-mean temperature equal to 2 K km^{-1}. Zonal and time averages are then taken.

Rates of change are calculated relative to surface air temperature and are corrected for the effects of finite temperature changes using equation (2) prior to multi-model averaging. The surface air temperature and surface specific humidity are generally reported in the model archive, with the exception of three models (GFDL-CM2.0, GFDL-CM2.1, MPI-ECHAM5) for which values at the lowest reported model level above the surface were used. The surface relative humidity shown in figure 2 is the relative humidity at the lowest reported model level above the surface.

4. Column water vapor

The multi-model mean rate of change of zonal-mean column water vapor is shown in figure 1. When plotted with respect to zonal-mean surface air temperature (left panel of figure 1), it ranges in value from less than 6% K^{-1} at high northern latitudes, to 8.6% K^{-1} at the equator, and has a maximum of 12% K^{-1} at 55°S. The local maximum at 55°S is related to smaller rates of surface warming in the southern ocean region (Meehl *et al* 2007), and becomes a local minimum when the rate of change is plotted with respect to global-mean surface air temperature (right panel of figure 1), although the range of values attained globally remains similar (6–12% K^{-1}).

Insight into the pattern of changes in the amount of water vapor can be gained by comparing with the changes in saturation column water vapor. To the extent that these are similar, the pattern of change may be purely an expression of changes in the thermal structure of the troposphere. The rate of change of saturation column water vapor is indeed similar to that of column water vapor, but it is greater in the subtropics and lower mid-latitudes, and smaller in the deep tropics, with differences of order 1–2% K^{-1} (figure 1).

To see whether these differences primarily result from changes in relative humidity or from vertical variations in the mean relative humidity, we also plot the rates of change corresponding to an invariant monthly-mean relative humidity distribution. These are generally smaller than the rates of change of saturation column water vapor, but still differ from those in column water vapor (figure 1). The implied decreases in mean relative humidity at roughly 10°–50° latitude in both hemispheres are consistent with previously reported decreases in free-tropospheric relative humidity at similar latitudes in simulations of global warming (Mitchell and Ingram 1992, Lorenz and DeWeaver 2007, Sherwood *et al* 2010, see also the top panels of figure 2). A temperature-of-last-saturation analysis suggests that subtropical decreases in free-tropospheric relative humidity in climate model simulations of global warming are primarily related to changes in the circulation (Wright *et al* 2010). It has also been argued that cross-isentropic fluxes of water vapor associated with convection are important for the control of free-tropospheric subtropical humidity (Schneider *et al* 2006, Couhert *et al* 2010). There is some observational evidence for negative trends in upper-tropospheric relative humidity at these latitudes (Bates and Jackson 2001), consistent with the pattern found in global warming simulations which extends quite deeply through the troposphere. Modeled changes in lower-tropospheric relative humidity are less robust in the deep tropics (Sherwood *et al* 2010).

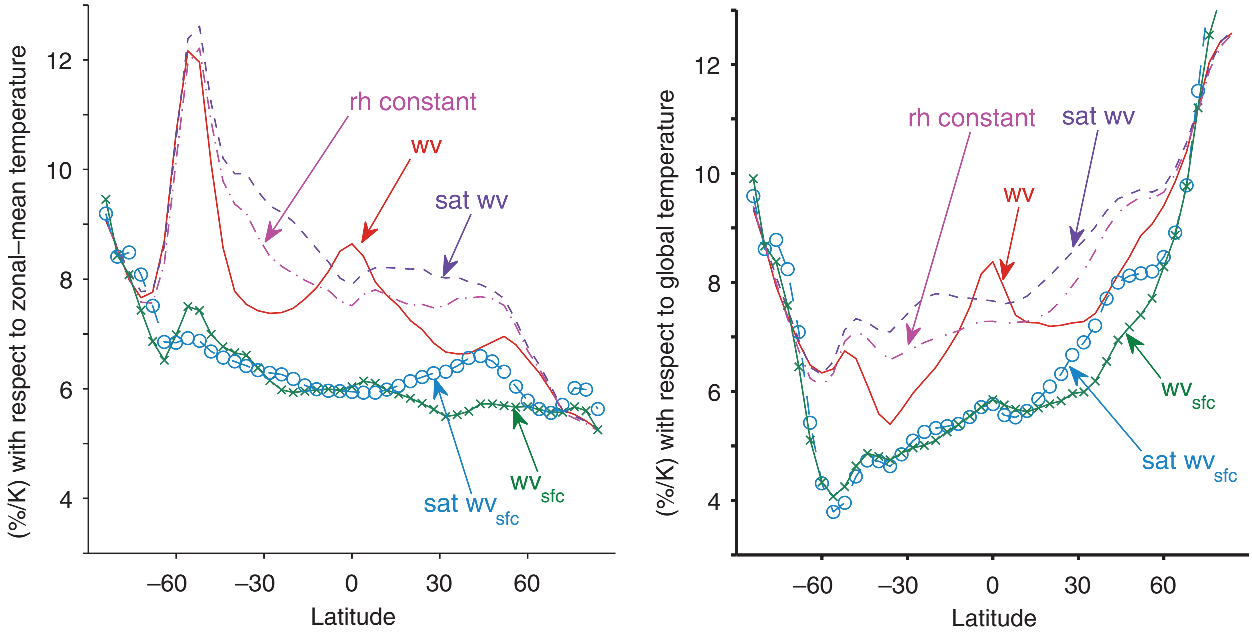

Figure 1. Rates of change (% K^{-1}) of column water vapor (red solid), column water vapor with an invariant distribution of relative humidity (pink dashed–dotted), saturation column water vapor (purple dashed), surface specific humidity (green line and crosses), and surface saturation specific humidity (blue line and circles). Rates of change are with respect to zonal-mean surface air temperature (left panel) and global-mean surface air temperature (right panel). The values shown are multi-model means of estimates of the differential rates of change based on equation (2) and differences between 1980–1999 and 2080–2099.

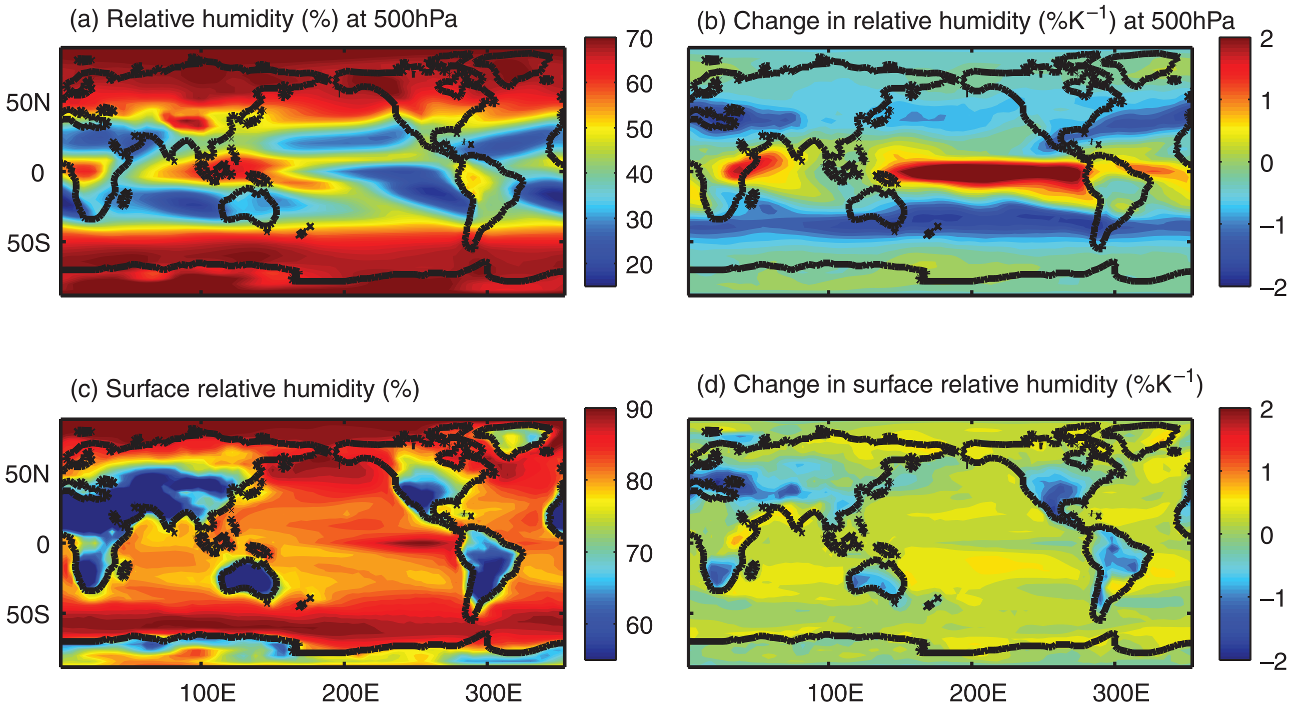

Figure 2. Mean relative humidity (%) for the period 1980–1999 and its rates of change (% K^{-1}) under climate change at 500 hPa ((a), (b)) and at the surface ((c), (d)). The rates of change are based on absolute rather than fractional changes in relative humidity (same time periods as in figure 1), are calculated with respect to global-mean surface air temperature, and are *not* modified using equation (2). Multi-model means are shown in all cases.

The global-mean rates of change with respect to global-mean surface temperature are 7.3% K^{-1} for column water vapor and 7.4% K^{-1} for column water vapor with invariant relative humidity (table 1). This implies that changes in mean relative humidity have almost no effect on the rate of change of global water vapor, although, as shown in figure 1, their impact is of order 1% K^{-1} at many latitudes. Interestingly, observed interannual variations in free-tropospheric relative

Table 1. Multi-model mean, minimum and maximum of rates of change of globally averaged quantities (all in % K^{-1} with respect to global-mean surface air temperature).

	Change in global-mean quantities (% K^{-1})		
	Mean	Minimum	Maximum
Column water vapor	7.3	6.5	8.2
Constant relative humidity	7.4	6.3	8.5
Saturation column water vapor	7.9	6.6	9.3
Surface specific humidity	5.7	5.2	6.2
Surface saturation specific humidity	5.9	5.2	6.4

humidity also tend to have regions of opposing changes so that the global-mean variability is muted (Dessler *et al* 2008).

The value of $r = 7.3\%\ K^{-1}$ for global-mean water vapor (which has been corrected for the finite temperature change as discussed in section 2) is roughly consistent with the value of $r_\Delta = 7.5\%\ K^{-1}$ found by Held and Soden (2006) based on two sets of simulations with global-mean surface air temperature changes of $\sim$0.6 K (for which $r = 7.3\%\ K^{-1}$ corresponds to $r_\Delta = 7.5\%\ K^{-1}$), and $\sim$2.5 K (for which $r = 7.3\%\ K^{-1}$ corresponds to $r_\Delta = 8.0\%\ K^{-1}$). The minimum and maximum rates of change of global water vapor over the range of models (table 1) indicate a model scatter of 1.7% K^{-1}, but the smaller fractional rate of change of column water vapor compared with Clausius–Clapeyron scaling in the subtropics and mid-latitudes is consistently found in individual model results (not shown).

5. Surface specific humidity

We also analyze the rate of change of surface specific humidity under climate change. This cannot be drastically different from the rate of change of column water vapor since water vapor is mostly concentrated near the surface (e.g., Schneider *et al* 2010). However, it may not behave in exactly the same way given vertical variations in mean temperature and relative humidity and their changes under global warming.

The rates of change of surface specific humidity are generally smaller than those in column water vapor except at high latitudes in both hemispheres (figure 1). This is also the case for surface saturation specific humidity, and so is partly related to the thermal structure of the atmosphere and its changes under climate change. The difference between the rates of change of column water vapor with invariant relative humidity and surface saturation specific humidity is of order 2% K^{-1} for latitudes in the range 50°S–50°N (figure 1). Thus, Clausius–Clapeyron scaling implies somewhat different rates of change for surface and column quantities.

The fractional changes in zonal-mean surface specific humidity and saturation specific humidity are very similar except in northern mid-latitudes (global rates of change of 5.7% K^{-1} and 5.9% K^{-1}, respectively; table 1). The deviations from Clausius–Clapeyron scaling in the northern hemisphere are suggestive of a difference in behavior over land and ocean. Figure 2 shows the changes in mean relative humidity versus latitude and longitude near the surface and at 500 hPa. The decreases in surface relative humidity occur primarily over continental interiors and are distinct from the more zonally banded changes at 500 hPa. A land/ocean contrast in the response of the surface relative humidity is not surprising given that the surface relative humidity over ocean is strongly constrained by the surface energy budget (Held and Soden 2000, Schneider *et al* 2010). Using a simplified surface energy budget, Schneider *et al* (2010) estimated a rate of increase of order 1% K^{-1} in surface relative humidity over ocean, consistent in order of magnitude with the changes over ocean shown in figure 2(d).[4] Smaller increases in surface relative humidity are implied if the surface winds or surface air temperature difference decrease in magnitude (cf Richter and Xie 2008). The simplified surface energy budget argument gives a rate of change of surface relative humidity that is proportional to 100% minus the mean surface relative humidity, seemingly consistent with the global pattern of greater changes over arid regions (figure 2(d)), but this argument does not apply over land because of the effects of limited moisture availability on evaporation rates over land.

The decreases in surface relative humidity over land may be related to the amplification of surface air temperature changes over land compared with over ocean[5]. This amplification is a feature of both transient and equilibrium climate change experiments (e.g., Manabe *et al* 1991, Meehl *et al* 2007, Joshi *et al* 2008), and is only partly related to the thermal inertia of the ocean in transient experiments. Assuming that the boundary-layer specific humidity is spatially homogenized to some extent by the circulation (e.g., a combination of mean zonal advection and vertical transports) and that the boundary-layer relative humidity over ocean remains constant, then the greater surface warming over land than ocean implies a decrease in boundary-layer relative humidity over land. We can make a rough estimate of the expected magnitude of decrease in relative humidity over land by considering the limit in which the specific humidity and its changes are the same over land and ocean. If the ratio of land to ocean surface warming is 1.5 (Joshi *et al* 2008), the rate of change of surface specific humidity over ocean is 6% K^{-1} (table 1), and the rate of change of surface saturation specific humidity over land is also 6% K^{-1} (table 1), then the *fractional* rate of decrease in relative humidity over land is $\sim$3% K^{-1} with respect to surface temperature over ocean. For a representative

[4] We generally refer to absolute rather than fractional rates of change in relative humidity.

[5] Joshi *et al* (2008) argue that boundary-layer relative humidity over land would decrease under global warming because of the nonlinearity of the Clausius–Clapeyron relation, even in the absence of a land/ocean contrast in surface warming. But their conceptual model does not support this conclusion if saturation specific humidities in the boundary layer and at the level of horizontal moisture convergence increase at roughly the same fractional rate (cf equation (3) of their paper).

value of surface relative humidity over land of 70%, this corresponds to a rate of decrease in surface relative humidity of $\sim$2% K^{-1}, which is comparable to the magnitude of decreases shown in figure 2(d). The circulation does not succeed in spatially homogenizing the specific humidity between land and ocean, and so a more detailed analysis is necessary to determine the exact factors contributing to the decrease in surface relative humidity over land.

Model simulations also indicate decreases in soil moisture under global warming in many of the regions where there is multi-model agreement (Wang 2005, Meehl *et al* 2007), but the changes are sensitive to the particular model, season, and region in question. Observational trends in surface relative humidity over land have been found to be statistically insignificant for the period 1976–2004 (Dai 2006). A more recent observational study found a large negative excursion in surface relative humidity over land for the decade from 1999–2008, and this was also postulated to be related to land/ocean contrast in surface temperature variability (Simmons *et al* 2010). If the modeled decreases in surface relative humidity over land are physically robust, then they can be expected to play an important role in the response to climate change of the surface energy budget, surface temperature, and hydrological cycle over land.

6. Conclusions

We have directly evaluated the contributions of several factors to the calculated rates of change in the amount of atmospheric water vapor under climate change. Our primary conclusions are as follows.

(i) Clausius–Clapeyron scaling under global warming is associated with a global rate of change of $\sim$7.4% K^{-1} for column water vapor, and $\sim$5.9% K^{-1} for surface specific humidity (table 1). But figure 1 shows there is a strong dependence on latitude and whether rates of change are expressed with respect to local or global-mean surface temperatures.

(ii) Deviations from Clausius–Clapeyron scaling of zonal-mean column water vapor result from decreases in relative humidity in the subtropics and mid-latitudes, and increases in the deep tropics.

(iii) Deviations from Clausius–Clapeyron scaling of surface specific humidity result from decreases in surface relative humidity over land that may be related to the amplification of surface warming over land compared with ocean.

(iv) The rate of change in the amount of water vapor is larger if calculated over a finite temperature change because of the quasi-exponential dependence of specific humidity on temperature (for example, by $\sim$2% K^{-1} for a temperature change of 6 K). Use of equation (2) allows comparison of rates of change in the amount of water vapor from different climate change simulations with different degrees of warming.

Precipitation intensity and precipitation extremes are sometimes assumed to scale with surface or column water vapor under climate change. According to our results, the difference between precipitation rates scaling with surface and column water vapor can be substantial: using column water vapor (8.4% K^{-1} at the equator) rather than surface specific humidity (5.8% K^{-1} at the equator) leads to a fractional rate of change that is 1.45 times greater at the equator (figure 1). In fact, the general dependence of cloud liquid water amounts and precipitation rates on temperature need not be the same as that for either surface specific humidity or column water vapor (Iribarne and Godson 1981, Betts and Harshvardhan 1987, O'Gorman and Schneider 2009a, 2009b, Schneider *et al* 2010). The dependence of the condensation rate on temperature is through the thermodynamic function $\mathrm{d}q_s/\mathrm{d}p|_{\theta^*}$, which is the derivative of the saturation specific humidity (q_s) with respect to pressure (p) at constant saturation equivalent potential temperature (θ^*), and which does not generally scale like q_s. Precipitation rates can be expected to scale with a vertical integral of $\mathrm{d}q_s/\mathrm{d}p|_{\theta^*}$ times the vertical pressure velocity (Iribarne and Godson 1981, O'Gorman and Schneider 2009a, 2009b). In the particular case of moist-adiabatic temperature lapse rates, sufficiently deep convection, and neglecting variations in the vertical velocity with height, this will roughly correspond to scaling with surface specific humidity (for example, in the case of tropical precipitation extremes). There is no support from these arguments for the scaling of precipitation rates with column water vapor under climate change, although free-tropospheric relative humidity does modulate tropical precipitation (e.g., Holloway and Neelin 2009, Muller *et al* 2009). Thus, although the rates of change in the amount of surface and column water vapor are useful to know for a number of reasons, related quantities such as precipitation rates and cloud liquid water amounts may have a different thermodynamic dependence.

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