Influence of condensate evaporation on water vapor and its stable isotopes in a GCM

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[1] The direct effect of condensate evaporation on atmospheric water vapor and its isotopic composition is assessed in a climate model. The model contains two parallel hydrologic cycles, an active one which influences the model physics and dynamics and a passive one which does not. Two model simulations are performed, one in which passive cloud and precipitation can evaporate and one in which they cannot. The active hydrologic cycles, and thus the simulated circulations and temperatures, are identical in both simulations. Eliminating passive condensate evaporation reduces the specific humidity in the passive cycle by around 5%; this reduction varies from a few percent to 25% of the control value, depending on location. Zonal mean water vapor in the lower and middle troposphere is enriched in HDO relative to the control case, and is depleted in the upper troposphere. **Citation:** Wright, J. S., A. H. Sobel, and G. A. Schmidt (2009), Influence of condensate evaporation on water vapor and its stable isotopes in a GCM, *Geophys. Res. Lett.*, *36*, L12804, doi:10.1029/2009GL038091.

1. Introduction

[2] Large-scale advection and condensation are sufficient to qualitatively determine the distribution of water vapor in the free troposphere [e.g., *Sherwood*, 1996; *Pierrehumbert and Roca*, 1998; *Galewsky et al.*, 2005; *Dessler and Minschwaner*, 2007]. These studies have relied on a class of simple models, termed 'advection-condensation' or 'large-scale control' (LSC) models. LSC models operate under the premise that detailed microphysics is unnecessary to simulate the large-scale distribution of water vapor. Whenever the local relative humidity (RH) exceeds a threshold value (typically between 90% and 100%), excess water vapor condenses and immediately falls out as precipitation.

[3] The qualitative success of LSC model simulations of water vapor implies that the influence of condensate evaporation on atmospheric humidity is small. There are unexplained quantitative discrepancies, however, which may be attributable in part to condensate evaporation. The magnitude of this contribution cannot be inferred in a straightforward way, as there are a number of errors and approximations in the LSC calculations which are difficult to assess precisely. These include resolution, errors in the winds and temperatures used to drive the model, the RH threshold at which condensation occurs, neglect of atmospheric mixing, and uncertainties in the observations of humidity to which the results are compared.

[4] In this study we sidestep this difficulty in a perfect model framework. We use a climate model (GCM) which contains a passive hydrologic cycle that does not influence the model physics or dynamics, in addition to the active one that does. We compare two simulations, one in which condensate is allowed to evaporate in the passive hydrologic cycle (CTL) and one in which it is assumed to be immediately removed as precipitation (LSC). The active hydrologic cycle remains unchanged; as a result, the two simulations are identical in every way other than the passive hydrologic cycle. This allows us to isolate the direct effects of condensate evaporation on atmospheric water vapor, excluding feedbacks that change the model circulation or temperature. Any difference in the passive water variables can be attributed to the direct effect of condensate evaporation. The influence of condensate evaporation on the circulation is of interest in its own right [e.g., *Bacmeister et al.*, 2006] but is not addressed here.

[5] Although we use the phrase 'large-scale condensation' to indicate the elimination of condensate from the model, our calculation does not necessarily require saturation at the grid scale for condensation to occur, but also uses the sub-grid scale convective parameterization. This makes it different from the true advection-condensation calculations done in some previous studies [e.g., *Dessler and Minschwaner*, 2007, and references therein].

[6] The contribution of condensate evaporation to the atmospheric water vapor distribution may also be constrained using models and observations of water vapor isotopes [*Moyer et al.*, 1996; *Gettelman and Webster*, 2005; *Schmidt et al.*, 2005; *Worden et al.*, 2007]. Heavy isotopes of water (e.g., HDO and $H_2^{18}O$) have a slightly higher vapor pressure than the most common isotope ($H_2^{16}O$), and preferentially exist in the condensed phase. Atmospheric water vapor is thus depleted in heavy isotopes relative to its evaporative source, and becomes progressively more depleted whenever condensation occurs. Evaporation of condensate may either deplete or enrich the isotopic content of surrounding vapor, depending on the relative isotopic content of the condensate and whether the evaporation is partial or complete (i.e., whether fractionation occurs – no fractionation occurs during complete evaporation). The isotopic ratios of atmospheric vapor thus represent an integrated history of water phase changes.

[7] Stable water isotopes are implemented in the GCM we use as a component of the passive hydrologic cycle [*Schmidt et al.*, 2005]. The elimination of passive condensate evaporation affects the isotopic ratios of atmospheric water vapor. We also prohibit both supersaturation with

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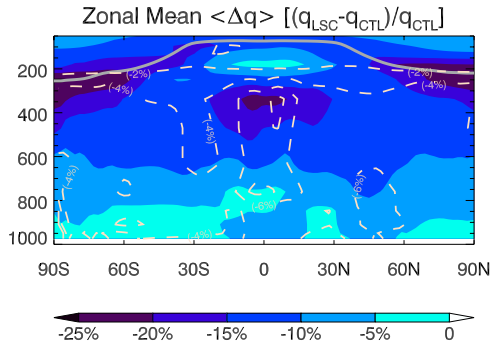

Figure 1. Zonal mean relative water vapor change between the CTL and LSC simulations. Absolute changes in relative humidity are shown as light dashed contours, with a contour interval of 2% RH. The zonal mean WMO tropopause is shown as a solid grey line.

respect to ice and isotopic equilibration between liquid droplets and surrounding vapor. The elimination of both supersaturation and condensate evaporation means that all water phase changes in the atmosphere occur at thermodynamic equilibrium. The hydrologic cycle in the LSC GCM is effectively a Rayleigh distillation model; that is, condensate is formed at isotopic equilibrium when RH is 100% (though again this need not be the case at the grid scale), and precipitation is immediately removed [cf. *Dansgaard*, 1964].

2. Model Formulation

[8] The GCM simulations are conducted using the Goddard Institute for Space Studies (GISS) ModelE [*Schmidt et al.*, 2006]. The model is run at 2° latitude by 2.5° longitude resolution with 20 vertical levels. Advection is computed using a quadratic upstream scheme [*Prather*, 1986], which yields an effective tracer resolution of approximately 0.7° × 0.8°. Sea surface temperatures and sea ice extent follow a fixed annual cycle averaged over 1975 to 1984, and all other boundary conditions are set to 1979 values as described by *Schmidt et al.* [2006]. The simulations are run for six years, of which the last five are analyzed.

[9] Stable isotopes of water, including $H_2^{16}O$, are incorporated into the model as passive tracers. This passive hydrologic cycle is implemented parallel to the GCM's active hydrologic cycle, and is subject to all of the same physics (advection, condensation, evaporation, diffusion, etc.). Isotope fractionation during phase changes generally occurs at thermal equilibrium, except for three cases: evaporation from the surface, deposition to ice in supersaturated conditions, and evaporation of liquid raindrops in unsaturated air [*Schmidt et al.*, 2005]. Liquid droplets equilibrate with the surrounding vapor. HDO abundance is expressed as a deviation from standard mean ocean water (SMOW), in units of per mil (‰):

$$\delta D = \left(\frac{R - R_{SMOW}}{R_{SMOW}}\right) \times 1000 \qquad (1)$$

where R is the ratio of the abundance of HDO to the abundance of water vapor. The abundance of $H_2^{18}O$ is expressed relative to δD, using the deuterium excess (d-excess = $\delta D - 8\delta^{18}O$) as described by *Dansgaard* [1964].

[10] We conduct and analyze two simulations. In the first, designated CTL, passive water vapor is pegged exactly to water vapor in the active hydrologic cycle. This simulation is nearly identical to the NEW simulation described by *Schmidt et al.* [2005], with two major differences. The first is that CTL is run at 2° × 2.5° horizontal resolution, whereas NEW was run at 4° × 5°. The second and more important difference is that in NEW, passive tracer amounts in the ocean are fixed and all other surface reservoirs of passive tracer are free to evolve; in CTL, all surface reservoirs of passive tracer are expressed as fixed ratios of active water. For lakes and groundwater we use SMOW; that is, the ratios are 1 for H_2O, 2.228×10^{-3} for $H_2^{18}O$, and 3.29×10^{-4} for HDO. The ratios are 2.172×10^{-3} (−25‰) for $H_2^{18}O$ and 2.639×10^{-4} (−198‰) for HDO in land ice, and 2.236×10^{-3} (4‰) for $H_2^{18}O$ and 3.358×10^{-4} (21‰) for HDO in sea ice. This modification does not affect passive water, but it does alter the evaporative fluxes of HDO and $H_2^{18}O$.

[11] The second simulation is designated LSC. The primary modification from CTL is that passive condensate is not allowed to evaporate. This change is effected in both the convective and large-scale (stratiform) cloud parameterizations, and includes evaporation of precipitation, cloud water, and condensate in unsaturated convective downdrafts. Tracer in condensate that would have evaporated is deleted instead; although this removes tracer from the system, it does not lead to long-term drift in the hydrologic cycle because the surface tracer reservoirs are fixed and provide an additional source. The model formulation is also adjusted so that passive tracer condensation occurs according to the passive RH rather than the active RH. Finally, supersaturation with respect to ice is prohibited in the passive hydrologic cycle; since rain evaporation in subsaturated air is also eliminated, this means that no kinetic (non-equilibrium) fractionation occurs during atmospheric phase changes, effectively rendering the LSC GCM a Rayleigh model. Elimination of supersaturation is only important to the stable isotopes, having a negligible effect on total water vapor. Supersaturation in the control case is parameterized as a function of temperature [*Schmidt et al.*, 2005].

3. Results

[12] Figure 1 shows the difference in zonal mean specific humidity between the LSC and CTL simulations, normalized by the CTL specific humidity. Eliminating condensate evaporation dries the modeled atmosphere everywhere by up to 25.7%. The largest drying occurs in the tropical upper troposphere (UT) between 500 hPa and 300 hPa and near the extratropical tropopause. The smallest drying occurs in the boundary layer and at the base of the tropical tropopause layer (TTL). The water vapor differences are normalized to remove the temperature dependence of water vapor and to better represent its radiative influence, since the longwave absorptivity of water vapor is logarithmic with concentration [*Soden*, 2005]. The zonal mean relative humidity, also

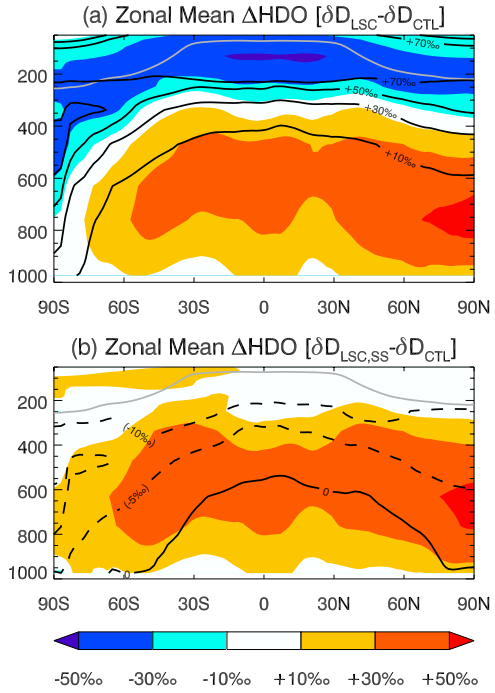

Figure 2. (a) Zonal mean change in δD (solid contours) and d-excess (black contours) between the CTL and LSC simulations. The contour interval for d-excess is 20‰. (b) As in Figure 2a but for a variation of the LSC simulation in which non-equilibrium fractionation due to supersaturation with respect to ice is allowed.

shown, decreases by amounts of order 5% RH (see Figure S1 for CTL climatology).[1] Decreases are slightly larger (6% to 8%) in the in the tropical deep convective zone and slightly smaller (~2%) in the tropical tropopause region.

[13] Corresponding zonal mean changes in δD and d-excess are shown in Figure 2a (see Figure S2 for CTL climatology). Water vapor is generally enriched by 10‰ to 50‰ in the lower and middle troposphere (surface to 400 hPa), with a maximum enrichment in the middle troposphere over the north pole of 53‰. Water vapor above 400 hPa is depleted by a similar amount, with a maximum decrease in δD of 54‰ in the TTL. The d-excess increases everywhere, with a particularly strong signal in the TTL. Figure 2b also shows zonal mean changes in δD and d-excess, but for a variation of the LSC simulation in which kinetic fractionation due to supersaturation with respect to ice is allowed. The associated deviations in δD from the CTL values are nearly identical to those shown in Figure 2a through most of the troposphere. In the upper troposphere, however, allowing kinetic fractionation during ice sedimentation replaces the strong depletion shown in Figure 2a with slight enrichment (~10‰) or near-zero change. The d-excess signal is also substantially different, with slight increases (less than 5‰) near the surface and slight decreases (5‰ to 10‰) aloft.

[14] Figure 3 shows the change in total column water vapor (see Figure S3 for CTL climatology) and mass-weighted column-average δD. The magnitude of the decrease is between 3.1% and 27.2% for water vapor path, similar to Figure 1. The largest drying occurs over the continents, and the smallest drying occurs over the subtropical oceans. Mass-weighted column-average δD increases nearly everywhere, except for over Antarctica. The maximum decrease is 25‰ over eastern Antarctica, and the maximum increase is 70‰ over subtropical South America.

4. Discussion

[15] The normalized change in water vapor amount shown in Figure 1 is relatively uniform in the free troposphere, particularly outside the tropics. This uniformity is also observed when the two GCM simulations are compared at individual levels (not shown), and is consistent with the qualitative agreement between LSC simulations and observations.

[16] The influence of condensate evaporation on water vapor amount in the GCM increases with altitude, reflecting the relative importance of condensate evaporation compared to surface evaporation. At low altitudes, evaporation from the surface is the primary vapor source; the relative importance of condensate evaporation grows as altitude increases. A similar relationship is observed in Figure 3. Total column water vapor is least affected by condensate evaporation over the subtropical oceans, where surface evaporation is high (5 mm d^{-1} to 6 mm d^{-1}). Conversely, the column dries most substantially over continental regions, particularly those where surface evaporation is small (less than 2 mm d^{-1}).

[17] The strongest drying due to eliminating condensate evaporation occurs in the tropical UT. This is due to the elimination of local condensate evaporation, particularly condensate detrained from deep convection. Previous LSC models do not simulate a strong decrease in this location; in fact, LSC simulations typically overpredict UT humidity in tropical convective regions [cf. *Galewsky et al.*, 2005; *Dessler and Minschwaner*, 2007]. The strong signal in the GCM should be viewed with caution, as this version of the ModelE is known to contain higher ice water paths than either observations or other GCMs [*Waliser et al.*, 2009]. The convective parameterization may detrain excessive

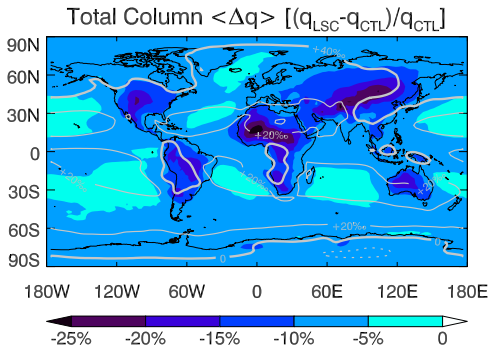

Figure 3. Relative change in total column water vapor (solid contours) and change in column average δD (grey contours) between the CTL and LSC simulations. The contour interval for δD is 20‰, ranging from −20‰ to 40‰. Negative changes are shown as dashed contours.

[1] Auxiliary materials are available in the HTML. doi:10.1029/2009GL038091.

condensate, resulting in excessive drying when condensate evaporation is disabled.

[18] Strong drying also occurs near the extratropical tropopause. This signal is not associated with local condensate evaporation. Analysis of the poleward and vertical transport of the water vapor tracer (not shown) reveals decreases at these levels of between 10% and 30% by mass. Most of this change is linked to reduced transport by mid-latitude eddies.

[19] Zonal mean water vapor in the GCM is generally enriched in heavy isotopes when condensate evaporation is eliminated (Figure 2). This enrichment occurs for three reasons. First, the evaporation of relatively depleted condensate, such as raindrops in the lower troposphere, is eliminated. This type of response occurs when the condensate that would have evaporated was formed at higher levels, where the vapor is typically more depleted. Second, the partial evaporation of condensate is eliminated. Even if the condensate is relatively enriched compared to the surrounding vapor, the preference of the heavy isotopes to remain in the condensed state leads to fractionation and potential depletion of vapor during partial evaporation. This mechanism is particularly relevant in a GCM, where the isotopic content of condensate must be parameterized as a property of the grid box or convective plume, rather than a property of individual hydrometeors. Third, the occurrence of condensation is reduced because of the decrease in RH. Condensation preferentially removes heavy isotopes and acts to deplete the vapor.

[20] If only the results of eliminating condensate evaporation are considered (i.e., supersaturation and kinetic effects are retained) then water vapor is enriched in heavy isotopes throughout the troposphere (Figure 2b), although this enrichment remains less pronounced in the lower and upper troposphere. In the lower troposphere the weakened response is because the signal is largely controlled by surface evaporation. In the UT it is due to the elimination of evaporation of lofted condensate. For the purposes of this discussion, we will consider lofted condensate to be condensate detrained by convection above 200 hPa. Although convective overshoot is not explicitly included in this version of the GCM, convection may detrain at levels as high as 110 hPa [*Schmidt et al.*, 2005]. Lofted condensate is generally enriched relative to the surrounding vapor, and has been invoked to explain the relative enrichment of water vapor isotopes in the TTL and lower stratosphere ($\sim -650‰$) compared to that expected from a Rayleigh distillation model ($-800‰$ to $-900‰$) [e.g., *Moyer et al.*, 1996; *Webster and Heymsfield*, 2003; *Dessler et al.*, 2007]. When condensate evaporation is eliminated in the GCM, the depletion of δD caused by eliminating evaporation of lofted condensate is effectively canceled by enrichment due to the reduced occurrence of condensation (Figure 2b).

[21] In the LSC model, we have also prohibited supersaturation with respect to ice. Deposition of vapor to ice occurs very rapidly in supersaturated conditions, which reduces the effectiveness of isotopic fractionation. When compared to deposition that occurs at thermodynamic equilibrium, this reduced effectiveness leaves the remaining vapor enriched in HDO relative to $H_2^{16}O$ and in $H_2^{18}O$ relative to HDO. The elimination of this process in the model thus results in a depletion of δD and an increase of d-excess in the UT and over Antarctica (Figure 2). The importance of this type of kinetic fractionation in the atmosphere remains poorly constrained, although it has been observed in laboratory experiments and its inclusion in models improves the simulation of isotopic ratios in Antarctic snow [*Jouzel and Merlivat*, 1984; *Schmidt et al.*, 2005] and in the TTL [*Gettelman and Webster*, 2005].

5. Conclusions

[22] We have performed a calculation in which condensate evaporation is eliminated in the passive water cycle without altering the circulation or temperature, and compared it to a control simulation in which condensate evaporation is present. Disabling condensate evaporation dries the modeled atmosphere by about 5% to 25%. In this model, that is the magnitude of the direct importance of microphysics to atmospheric humidity.

[23] Decreases of specific humidity are most pronounced in the tropical UT, at the extratropical tropopause, and over land. The first is related to evaporation of detrained condensate in convective anvils, the second is due to non-local condensate evaporation and transport, and the third is due to the heightened importance of rain re-evaporation in these regions relative to surface evaporation. Humidity over the subtropical oceans and in the TTL is relatively insensitive to the elimination of condensate evaporation.

[24] Water vapor in the LSC simulation is enriched in heavy isotopes in the lower and middle troposphere and depleted in the UT relative to the CTL simulation. The enrichment of lower and middle tropospheric vapor occurs in part because evaporation of relatively depleted condensate (e.g., raindrops) is eliminated, in part because depletion due to partial evaporation of condensate is eliminated, and in part because condensation occurs less frequently due to the reduction in RH. The depletion of UT vapor is caused primarily by the prohibition of supersaturation with respect to ice and associated kinetic fractionation. Depletion caused by eliminating the sublimation of convectively detrained condensate also contributes, but is effectively offset by enrichment due to the reduced occurrence of condensation.

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References

Bacmeister, J. T., M. J. Suarez, and F. R. Robertson (2006), Rain reevaporation, boundary layer-convection interactions, and Pacific rainfall patterns in an AGCM, *J. Atmos. Sci.*, *63*, 3383–3403, doi:10.1175/JAS3791.1.

Dansgaard, W. (1964), Stable isotopes in precipitation, *Tellus*, *16*, 436–458.

Dessler, A. E., and K. Minschwaner (2007), An analysis of the regulation of tropical tropospheric water vapor, *J. Geophys. Res.*, *112*, D10120, doi:10.1029/2006JD007683.

Dessler, A. E., T. F. Hanisco, and S. Fueglistaler (2007), Effects of convective ice lofting on H_2O and HDO in the tropical tropopause layer, *J. Geophys. Res.*, *112*, D18309, doi:10.1029/2007JD008609.

Galewsky, J., A. Sobel, and I. Held (2005), Diagnosis of subtropical humidity dynamics using tracers of last saturation, *J. Atmos. Sci.*, *62*, 3353–3367.

Gettelman, A., and C. R. Webster (2005), Simulation of water isotope abundances in the upper troposphere and lower stratosphere and implications for stratosphere troposphere exchange, *J. Geophys. Res.*, *110*, D17301, doi:10.1029/2004JD004812.

Jouzel, J., and L. Merlivat (1984), Deuterium and oxygen 18 in precipitation: Modeling of the isotopic effects during snow formation, *J. Geophys. Res.*, *89*, 11,749–11,757.

Moyer, E. J., F. W. Irion, Y. L. Yung, and M. R. Gunson (1996), ATMOS stratospheric deuterated water and implications for troposphere-stratosphere transport, *Geophys. Res. Lett.*, *23*, 2385–2388.

Pierrehumbert, R. T., and R. Roca (1998), Evidence for control of Atlantic subtropical humidity by large scale advection, *Geophys. Res. Lett.*, *25*, 4537–4540.

Prather, M. J. (1986), Numerical advection by conservation of second order moments, *J. Geophys. Res.*, *91*, 6671–6680.

Schmidt, G. A., G. Hoffman, D. T. Shindell, and Y. Hu (2005), Modeling atmospheric stable water isotopes and the potential for constraining cloud processes and stratosphere-troposphere exchange, *J. Geophys. Res.*, *110*, D21314, doi:10.1029/2005JD005790.

Schmidt, G. A., et al. (2006), Present-day atmospheric simulations using GISS ModelE: Comparison to in situ, satellite, and reanalysis data, *J. Clim.*, *19*, 153–192, doi:10.1175/JCLI3612.1.

Sherwood, S. C. (1996), Maintenance of the free-tropospheric tropical water vapor distribution. Part II: Simulation by large-scale advection, *J. Clim.*, *9*, 2919–2934.

Soden, B. J. (2005), The radiative signature of upper tropospheric moistening, *Science*, *310*, 841–844.

Waliser, D. E., et al. (2009), Cloud ice: A climate model challenge with signs and expectations of progress, *J. Geophys. Res.*, *114*, D00A21, doi:10.1029/2008JD010015.

Webster, C. R., and A. J. Heymsfield (2003), Water isotope ratios D/H, $^{18}O/^{16}O$, $^{17}O/^{16}O$ in and out of clouds map dehydration pathways, *Science*, *302*, 1742–1745.

Worden, J., D. Noone, and K. Bowman (2007), Importance of rain evaporation and continental convection in the tropical water cycle, *Nature*, *445*, 528–532, doi:10.1038/nature05508.

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