Science AAAS

www.sciencemag.org/cgi/content/full/1115602/DC1

Supporting Online Material for

The Radiative Signature of Upper Tropospheric Moistening

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Published 6 October 2005 on *Science* Express
DOI: 10.1126/science.1115602

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Water vapor in the atmosphere's boundary layer (which extends from the surface to ~800 mb) is widely expected to exhibit a constant relative humidity behavior. A discussion of the physics underlying this tight coupling to Clausius-Clapeyron in the lower troposphere can be found in (S1), where it is argued that it is not simply the proximity of the lower troposphere to the saturated oceans that is responsible, but a more fundamental constraint which arises from the sensitivity of surface evaporation to changes in relative humidity. However, there are no such constraints on water vapor in the free troposphere (the portion of the troposphere above the boundary layer). It is the moisture response in this region which lies at the heart of the water vapor feedback debate and where convective and cloud microphysical processes have long been hypothesized to play an important role (S1).

1) High Resolution Infrared Sounder (HIRS) Radiance Observations

The HIRS channel 12 is sensitive to water vapor over a broad layer of the upper troposphere which shifts from ~200-500 mb in the tropics to ~350-700 mb in the mid-latitudes (S2), closely following the latitudinal variations in tropopause height. A strength of using radiances rather than a retrieved quantity on constant pressure levels is that the vertical sampling of the channel 12 weighting function closely follows the levels in the atmosphere to which water vapor feedback is most sensitive.

To homogenize HIRS measurements from different satellites, we compute the inter-satellite offsets by minimizing the differences of overlapped global, monthly mean radiances between satellites. In most cases, the standard error of the calibration offset between satellites is an order of magnitude smaller than the mean offset, indicating that there is a large enough overlap between satellites to obtain a stable intercalibration. One exception is the offset between NOAA-7 and NOAA-9 which have less than one month of overlap with each other. Here we apply a curve fitting to HIRS anomaly time series from September to December 1984 (NOAA-7 observations) and then extrapolate this curve to January 1985. Similar fitting is also applied to January to April 1985 (NOAA-9 observations) and then the curve is extrapolated back to December 1984. Then adjustment for NOAA-7 vs. NOAA-9 is obtained by minimizing the radiance differences of December 1984 and January 1985. Further details regarding the intercalibration procedure are provided in (S3).

Because the diurnal cycle of T12 is small, conclusions regarding the trends in T12 are not affected by the aliasing effects of drift in the observation times even for those satellites with the largest equatorial crossing time drift [S4-S5]. Furthermore, we have also sampled the model output according to the HIRS observation times for each NOAA satellite and find no significant change in either the GCM or “GCM no moistening” simulations of T12.

Similar trends are obtained if the intercalibrated T12 data set of Bates and Jackson (S6) is used. This data set ends in 1998 and only covers 70S to 70N. For this domain, the linear trend from 1982-1998 from their version of the T12 data is 0.02 K/decade, which is consistent with that obtained from our intercalibrated T12 (0.03 K/decade). It also compares favorably to that obtained from the GCM simulations (0.02 – 0.03 K/decade). In contrast, the GCM simulations of T12 with no moistening have a trend of 0.22 K/decade over this period.

2) Climate model radiance simulations

The “constant relative humidity” T12 was calculated by replacing the model-predicted water vapor mixing ratio with that obtained by computed from the model simulated temperature field using a prescribed, seasonally-varying climatology of relative humidity from the model. Thus as the atmosphere warms over the period 1982-2004, the mixing ratio used to compute the T12 increases at a constant relative humidity rate. These results are shown by the red dashed lines in Figures 2 and 3.

The “no moistening” T12 was calculated as above, but replacing the model-predicted water vapor mixing ratio with a prescribed, seasonally-varying profile which does not vary from one year to the next. Thus, as the atmosphere warms over the period 1982-2004, the mixing ratio used to compute the T12 remains unchanged. Because the change in T12 is proportional to the fractional change in water vapor mixing ratio, the trends in T2-T12 for any intermediate scenario can be estimated as a linear combination of the constant relative humidity and constant mixing ratio scenarios.

The climate model simulations of T2 are computed using the RTTOVS radiative transfer code (S7). Because T2 receives some contribution from stratospheric temperature, its anomalies are affected by stratospheric cooling trends which arise primarily from decreases in stratospheric ozone, as well as other natural and anthropogenic forcings. The climate model simulations described here do not include these stratospheric forcings and it is therefore expected that their simulated trend in T2 should be slightly larger than observed. Simulations for the period 1982-2000 which include these forcings (the ozone forcing data were not available after 2000) suggest that their absence introduces a positive bias in the model simulated T2 trend of ~0.02 K/decade.

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Figures

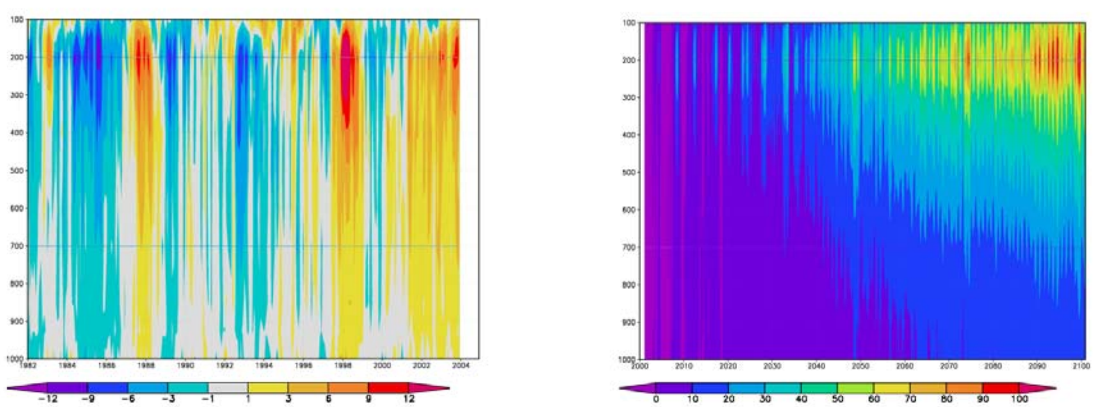

Figure S1: GCM simulations of the percentage change in water vapor mass as a function of height from atmospheric model simulations with observed SST for 1982-2004 (left); and from coupled ocean-atmosphere model simulations using projected increases in greenhouse gases over the period 2000-2100 (right).