Water Vapor Feedback:

From classic calculations such as Manabe and Wetherald(1967) we see that if relative humidity and lapse rate are fixed, water vapor feedback is very strong and gives a relatively sensitive climate. In chapter 9 of GPC the response factor to climate forcing is defined.

$$\frac{dT_s}{dQ} = \lambda_R \tag{1}$$

Where T_s is the surface temperature, usually assumed to be the global mean, and Q is an external energy forcing imposed in the climate, usually in Wm^{-2}. The sensitivity parameter thus measures the response of surface temperature to a climate forcing of a given magnitude. Water vapor feedback approximately doubles this feedback parameter, when relative humidity is assumed constant.

If the lapse rate is allowed to vary, but the relative humidity is assumed fixed, the effect on OLR of variations in lapse rate tend to be canceled by the effects of the implied changes in the vertical distribution of water vapor. If the lapse rate is less, this reduces the greenhouse effect because the emission temperature is closer to the surface temperature. If the warmer air at higher altitudes associated with decreased lapse rate also contains more water vapor (following an assumption of fixed relative humidity at all heights), then the emission level is moved upward to a higher altitude where the temperature is less. So with an assumption of fixed relative humidity at each level, the influence of lapse rate and humidity distribution effects on the greenhouse effect tend to cancel (e.g. Cess, 1975).

Hansen, et al (1984) proposed using a different feedback quantification methodology adopted from electronics. They define a feedback factor *f* from:

$$\Delta T_{eq} = f \; \Delta T_o, \tag{1}$$

where ΔT_{eq} is the temperature change produced, and ΔT_o is the temperature change that would have occurred in the absence of the feedback in question. We assume that the feedbacks are additive, so that we can write,

$$\Delta T_{eq} = \Delta T_o + \Delta T_{feedback} \tag{2}$$

and then define the gain as,

$$g = \frac{\Delta T_{feedback}}{\Delta T_{eq}} \tag{3}$$

With an assumption of fixed relative humidity, water vapor feedback gives a gain of 0.5 and a feedback factor of 2, so it doubles the temperature response over what you would get with fixed specific humidity and fixed lapse rate. Of course, we have no particular reason to have confidence that these assumptions are precisely accurate, and we need to investigate the sensitivity of our conclusions to them. There are a number of issues to address and several approaches to them, which we will discuss below. First let's use the gain/feedback methodology to explore the significance of having a single large feedback like the water vapor feedback. Once we have a strong feedback operating, adding smaller feedbacks has a surprisingly strong effect on the temperature response, which is the bottom line in this.

The gains are additive, so if we have two feedbacks working, the total gain is $g = g_1 + g_2$, but the feedback factors combine differently:

$$f = \frac{f_1 f_2}{f_1 + f_2 - f_1 f_2} \tag{4}$$

Suppose that the forcing is doubling of carbon dioxide, which produces a warming of 1˚C in the absence of feedbacks. Next add water vapor feedback, which we presume to have a gain of 0.5 and give us a total equilibrium temperature change of 2˚C. Next add a smaller feedback with a gain of 0.25. Using (4), the feedback factor with these two feedbacks is 4.0! We expect a temperature response of 4˚C!! Feedbacks amplify the temperature response to other feedbacks, so we need to look at all the little feedbacks in the system if we want to predict the temperature response with precision. Of course the strength of the strong feedbacks needs to be known with precision also.

1.) Sensitivity of the greenhouse effect to the vertical distribution of water vapor

We define the clear-sky greenhouse forcing as the difference between the longwave emission from the surface, and the outgoing longwave radiation. This is the amount by which the clear atmosphere decreases the escaping longwave emission.

Clear-Sky Greenhouse Forcing = Surface longwave emission - OLR (5)

One can also define a normalized greenhouse effect by dividing the above difference by the surface longwave emission.

Most of the water vapor in the atmosphere is quite close to the surface, so most of the infrared opacity is quite close to the surface. If we think of blackbody emission, the clear-sky greenhouse is related to the difference between the surface temperature and the emission temperature for clear skies. What does this depend on? We can increase this difference by a) increasing or elevating the infrared absorbers in the atmosphere, or b) increasing the lapse rate of temperature. Since water vapor is the primary longwave absorber in the atmosphere, we can look at how changes in it affect the surface temperature in a radiative-convective model, or how they affect the top-of-atmosphere flux in a pure radiative flux calculation.

Such calculations were done by Shine and Sinha(1991), largely in response to an assertion by Lindzen(1990) that warming would lead to drying of the upper troposphere, and that this would provide a negative feedback. They asked two questions: At what level would a fixed change in mass mixing ratio of water vapor provide the biggest surface temperature response in a 1-D radiative/convective model, and for what levels would a given fractional change (say 10%) of some standard profile give the biggest effect. In the first case, a fixed change of specific humidity of .001 g/kg in a 50mb thick layer gives the strongest response up near the tropopause (Fig. 1). This is because the temperatures are coldest there and the water mixing ratio is smallest, so the difference between the upward flux that is absorbed and the upward flux that is emitted by the added gas is maximized. In the second case, the mixing ratio was changed by 10% of its local value in 50mbar-thick layers. So rather than an absolute change in water vapor mass, we are considering a relative change of the local value. This produces much larger changes

[THERE WAS A PAGE OF FIGURES FROM SHINE AND SINHA HERE.]

in the mass mixing ratio of water vapor at lower levels, and the largest contributions to surface temperature come from around the 800mb level, especially in the tropics, although the contributions are large through most of the troposphere. 800mb is near the top of the mixed layer in the tropics. Much of the lower level contributions for tropical profiles comes from continuum absorption. in the window region 800-1200 cm^{-1}, or 8-12 microns(Fig. 4). In the tropics, the temperature response to a 10% change in humidity near the tropopause is small, because the saturation mixing ratio of water vapor there is so tiny.

2) Observational Approaches to Water Vapor Greenhouse feedback.

One can use the observed regional and seasonal variations in water vapor, temperature and radiative heating rates to get some observation insight into how things work. Raval and Ramanathan(1989) used surface temperatures and ERBE clear-sky estimates to estimate the clear-sky greenhouse effect. They found that it increased with surface temperature much in accord with the fixed relative humidity-fixed lapse rate models. Hallberg and Inamdar(1993) showed that the super-greenhouse effect, a more rapid than expected increase in the greenhouse effect at the highest temperatures was related to four causes:

- Increases in saturation humidity at high temperatures
- Increased continuum absorption for these high humidity, high temperature cases
- Increased water vapor in the upper troposphere associated with the presence of deep convection
- Increased lapse rate

It is also possible to use observations of temperature and humidity to calculate the relative contributions of temperature and humidity to OLR or greenhouse effect changes. A huge literature exists in this area. Generally it shows that seasonal variations in relative humidity and lapse rate are both important in determining the seasonal variations in clear-sky greenhouse effect (Webb et al. 1993, Bony and Duvel 1993, Sinha 1995, Bony et al, 1997). Clouds generally amplify these changes, since cloud occurs in association with high upper tropospheric humidities. Also, it is observed that there can be large cancellations between lapse rate and humidity distribution changes. For example, when the upper tropospheric temperature increases, the humidity usually also increases, and the effects of these changes on the greenhouse effect tend to cancel, as mentioned earlier in these notes.

3) One dimensional models with predicted lapse rate and humidity

Since we can see from calculation or observation that both the lapse rate and the vertical distribution of humidity play important roles in determining the clear-sky greenhouse effect, we need to have a theory that predicts the vertical distribution of temperature and humidity and their response to climate change. This is very difficult, since it involves the interaction of cloud-scale and large-scale processes, which is an unsolved problem. Some interesting possibilities can be explored by considering the model used by Sarachik(1978) and Sun and Lindzen(1993a).

Sun and Lindzen(1993b) consider the 1 dimensional energy balance of the tropics. The assumptions that are used in this simple parameterizations for convective/radiative interaction are 1) that the convective areas occupy a small fraction of the area of the tropics, so that the effect of the convection on the area averaged thermodynamics can be though of as the heating caused by the uniform subsidence, which is just equal and opposite to the convective mass flux M(z). The heat balance for the free troposphere is thus:

$$M\frac{ds}{dz} - R = 0 \tag{6}$$

which is to say that the downward advection of dry static energy s just balances the radiative cooling $R(z)$. In the stratosphere we have radiative equilibrium, so $R(z) = 0$.

If we integrate the subsidence heating over the free troposphere from the top of the PBL, z_b to the top of the troposphere, it must just balance the net radiative heat flux downward across the PBL top, assuming no net radiation at the top of the atmosphere.

$$\int_{z_b}^{z_t} M\frac{ds}{dz}\,dz = S(z_b) - F(z_b) \tag{7}$$

At the PBL top we have that the dry static energy flux across the PBL top must equal the latent heat flux going upward.

$$M(z_b)\,\Delta s \;=\; \alpha\, L\, E \tag{8}$$

where Δs is the jump of dry static energy across the PBL top, α is the entrainment ratio and E is the evaporation.

The heat budget of the PBL is then,

$$M(z_b)\,\Delta s \;+ SH \;=\; (S(0) - F(0)) - (S(z_b) - F(z_b)) \tag{9}$$

where S and F are the net downward solar flux and the net upward longwave flux, respectively. The surface heat budget is,

$$(S(0) - F(0)) \;=\; LE + SH \tag{10}$$

From (7), (9) and (10) we can derive,

$$\int_{z_b}^{z_t} M\frac{ds}{dz}\,dz + M(z_b)\,\Delta s = LE \tag{11}$$

We can also assume that the tropopause occurs where the dry static energy equals the moist static energy of the surface air (also the level of neutral buoyancy).

$$s(0) \;+\; RH\,L\,q^*(0) \;=\; s(z_t) \tag{12}$$

To make things easy, but not necessarily accurate, we can assume that $M(z)$ is a constant with height in the free troposphere. Then we can use (11) and (12) to solve for M,

$$M = \frac{E}{RH\, q^*(0)} \tag{13}$$

which means that the mass flux is set by the requirement to move water out of the PBL, or dry air in, equally. You can use the bulk aerodynamic formula approximation (9.37) in GPC to show that M is rather insensitive to surface temperature. if one assumes that relative humidity, wind speed and drag coefficient in the PBL are all insensitive to the temperature. To close the model fully, we can assume that the air-sea temperature difference needed for the bulk aerodynamic formulas is a fixed 1˚K. Now all we need is a radiative transfer code to calculate the radiation fluxes and we are in business.

Sun and Lindzen(1993b) used this model to investigate the problem of tropical SST and tropical mountain snowlines. They accepted both data points as accurate and set out to explain how the lapse rate could have changed so much during the last ice age. They found that if you *increase*(?) the humidity of the free troposphere during the last ice age, then you can cool down the free troposphere. You need about a 20% change in humidity to do it. This is at odds with lots of paleoclimatic evidence that suggests the tropics were drier during the ice age. Things such as dust and $\delta^{18}O$ in glaciers.

The Water Balance for this model:

Sun and Lindzen(1993b) also looked at the corresponding moisture model. Here again we have a balance between vertical advection and an internal source, this time the source comes from the convection.

$$M \frac{dq}{dz} + \rho_a E_r = 0 \tag{14}$$

Here q is the vapor mixing ratio, ρ_a is the air density, and the product $\rho_a E_r$ is the internal moisture source or the moistening effect of convection. We can parameterize this moisture source in the following way

$$\rho_a E_r = \rho_a \alpha (q^* - q) \tag{15}$$

where alpha is an exchange coefficient, not the same as the previous alpha used.

So we get,

$$M \frac{dq}{dz} + \alpha \rho_a (q^* - q) = 0 \tag{16}$$

We can rewrite this as an equation for the relative humidity RH,

$$\frac{d\,RH}{dz} - \frac{\mathrm{RH}}{\mathrm{h}_1} + \frac{1}{\mathrm{h}_2}\exp(-z/H) = 0 \tag{17}$$

Where H is the scale height, and h_1 and h_2 are parameters with dimensions of height.

$$\frac{1}{h_2} = \frac{\alpha \rho_o}{M} \tag{18}$$

$$\frac{1}{h_1} = \frac{1}{h_3} + \frac{1}{h_2} \exp(-z / H) \qquad (19)$$

$$\frac{1}{h_3} = -\frac{d \ln q^*}{dz} \qquad (20)$$

ρ_o is the density of the surface air. Because saturation humidity q^* decreases approximately exponentially with height, h_3 is nearly constant with height. Once this is accepted, then we see that *RH*, the relative humidity, is not explicitly dependent on temperature. Also, we had previously noted that *M* is almost independent of temperature, because the warming and cooling rate of the circulation stay in balance even when the absolute values go way up with the saturation vapor pressure. So unless there is some cloud physical process buried in h_2 that depends on temperature, then we should expect *RH* to keep about the same *RH* profile when the climate warms or cools. The key parameter then is α, which measures the rate at which humidity is exchanged between the cloud and its environment. Sun and Lindzen(1993b) argue that this parameter should vary inversely with temperature, such that a warmed climate will be drier and a cold climate wetter. They are led to this point of view by the CLIMAP SST and ice-age snowline data, that suggest larger lapse rates during the ice age. Thus they take as a starting point that the lapse rate increase with decreasing temperature. They then conclude that this would mean more CAPE at lower temperatures (Convective Available Potential Energy). Modern data indicate more hydrometeors when CAPE is higher, so they conclude that this will be more hydrometeors and higher α during a colder climate. Conversely during a warmer climate they expect the opposite, so that the free troposphere will dry out and give less greenhouse effect and a negative feedback. This line of reasoning is obviously tenuous, particularly given the strong evidence that the ice ages were drier in the tropics.

References:

Bony, S. and J.-P. Duvel, 1994: Influence of the vertical structure of the atmosphere on the seasonal variation of precipitable water and greenhouse effect. *J. Geophys. Res.*, **99**, 12,963-12,980.

Bony, S., K. M. Lau and Y. C. Sud, 1997: Sea surface temperature and large-scale circulation influences on tropical greenhouse effect and cloud radiative forcing. *Journal Of Climate*, **10**, 2055-2077.

Cess, R. D., 1975: Global Climate Change: An investigation of atmospheric feedback mechanisms. *Tellus*, **27**, 193-198.

Hallberg, R. and A. K. Inamdar, 1993: Observations of seasonal variations in atmospheric greenhouse trapping and its enhancement at high sea surface temperature. *J. Climate*, **6**, 920-930.

Hansen, J., A. Lacis, D. Rind, G. Russell, P. Stone, I. Fung, J. Lerner and R. Ruedy, 1984: Climate sensitivity: Analysis of feedback mechanisms. *Climate Processes and Climate Sensitivity*. 29, J. E. Hansen and T. Takahashi, Eds., AGU, 130-163.

Inamdar, A. K. and V. Ramanathan, 1994: Physics of greenhouse effect and convection in warm oceans. *Journal of Climate*, **7**, 715-31.

Lindzen, R. S., 1990: Some coolness concerning global warming. *Bull. Amer. Meteor. Soc.*, **71**, 1465-1467.

Raval, A. and V. Ramanathan, 1989: Observational determination of the greenhouse effect. *Nature*, **342**, 758-61.

Sarachik, E. S., 1978: Tropical sea-surface temperature: An interactive one-dimensional atmosphere ocean model. *Dyn. Atmos. Ocean*, **2**, 455-469.

Shine, K. P. and A. Sinha, 1991: Sensitivity of the Earth's climate to height-dependent changes in the water vapour mixing ratio. *Nature*, **354**, 382-4.

Sinha, A. and J. E. Harries, 1997: The Earth's clear-sky radiation budget and water vapor absorption in the far infrared. *Journal of Climate*, **10**, 1601-14.

Sinha, A., 1995: Relative influence of lapse rate and water vapor on the greenhouse effect. *J. Geophys. Res.*, **100**, 5095-5103.

Sun, D.-Z. and R. S. Lindzen, 1993a: Distribution of Tropical Tropospheric Water Vapor. *J. Atmos. Sci.*, **50**, 1643-1660.

Sun, D.-Z. and R. S. Lindzen, 1993b: Water vapor feedback and the ice age snowline record. *Ann. Geophysicae,* **11**, 204-215.

Webb, M. J., A. Slingo and G. L. Stephens, 1993: Seasonal variations of the clear-sky greenhouse effect: The role of changes in the atmospheric temperatures and humidities. *Clim. Dyn.*, **9**, 117-129.

Tropical Humidity Distribution and OLR:

The humidity distribution in the tropics varies greatly depending on whether convection has recently moistened the air column, or radiative cooling has enabled prior subsidence of the air column. On the following page two relative humidity profiles from the rainy season at Singapore and the dry season at San Jose, Puerto Rico are shown to indicate the general differences between these regimes(from Hartmann, et al., 1992). In the near vicinity of convection the mean air humidity is about 80% at all levels in the troposphere. In the dry zone, the humidity below the inversion is also high, but above the inversion the humidity falls off rapidly to very low values. In this plot it asymptotes to 20%, but that is an artifact of the radiosonde humidity measurement system, the actual humidities may be even drier at levels above the inversion.

From the calculations of Shine and Sinha(1991) shown previously, we know that the sensitivities of the OLR and radiative/convective equilibrium temperature to a 10% change in specific humidity peak near the tropical inversion, but are significant up to 300mb or so. If the boundary layer humidity and boundary layer depth are fixed, then the OLR and the surface temperature in equilibrium should be quite sensitive to the vertical distribution of humidity in the upper tropical troposphere, as Lindzen(1990) has utilized in his attempt to debunk positive relative humidty feedback. Pierrehumbert(1995) has called the dry zones of the tropics ‘radiator fins’ because the OLR tends to be high in these regions, and so they are regions where energy is effectively lost to space.

In the second figure on the following page, we show the sensitivity of the OLR to the humidity distribution. In the ‘moist’ case, the relative humidity is 80% at all levels. In the ‘dry’ case the relative humidity is 80% below 800mb and 20% above 800mb. Two temperature profiles are assumed. In the fixed tropical case, the tropical mean lapse rate is assumed at all levels. In the moist adiabatic lapse rate case the lapse rate follows a moist adiabat at all levels. The free parameter is the surface temperature, which becomes the independent variable of the graph. In the case of the moist adiabatic lapse rate(dotted lines) the lapse rate in the lower and middle troposphere decreases, so that the greenhouse effect should be reduced. For temperatures below about 305K, however, the difference between the fixed lapse rate and the moist adiabat is small. This is because the effect on OLR of the the warmer emission temperatures at upper levels is approximately cancelled by the fact that the emission level is raised by the increased specific humidity at upper levels that results when the temperture is raised, but the relative humidity is kept fixed.

If you look at 300K, the OLR is reduced from about 280 Wm^{-2} in the ‘dry’ case to about 250 Wm^{-2} in the ‘moist’ case. So upper tropospheric humidity differences are significant. If you let the upper troposphere get drier, the effect would be even larger. The blackbody emission at 300K is 459 Wm^{-2}, so the clear-sky greenhouse effect increaases from about 180 to about 210 Wm^{-2}, when you moisten the upper troposphere.

References:

Hartmann, D. L., M. E. Ockert-Bell and M. L. Michelsen, 1992: The effect of cloud type on Earth's energy balance: Global Analysis. *J. Climate*, **5**, 1281-1304.

Lindzen, R. S., 1990: Some coolness concerning global warming. *Bull. Amer. Meteor. Soc.*, **71**, 1465-1467.

Pierrehumbert, R. T., 1995: Thermostats, radiator fins and the local runaway greenhouse. *J. Atmos. Sci.*, **52**, 1784-1806.

Shine, K. P. and A. Sinha, 1991: Sensitivity of the Earth's climate to height-dependent changes in the water vapour mixing ratio. *Nature*, **354**, 382-4.