Ann. Geophysicae 11, 204–215 (1993) © EGS – Springer-Verlag 1993

Annales Geophysicae

Water vapor feedback and the ice age snowline record

De-Zheng Sun [1] **and Richard S. Lindzen** [2]

Center for Meteorology and Physical Oceanography, Building 54, Room 1720, Massachusetts Institute of Technology, Cambridge, MA 02139, USA

Received June 29, 1992; revised November 9, 1992; accepted November 30, 1992

Abstract. The CLIMAP reconstruction of the Earth's climate during the last major glaciation 18 000 years ago (ybp) found that while average tropical sea surface temperatures were no more than about 1 °C cooler than at present, snow lines on high peaks had descended 1 km. This suggests a temperature reduction of about 5 °C in the mid-troposphere. Since the tropical atmosphere (above the tradewind boundary layer) cannot sustain large horizontal temperature gradients, this feature is likely to be characteristic – implying the lapse rate in the lower half of the tropical troposphere during the last glaciation was about 20% greater than at present. We note that such a reduction in static stability requires a reduction of radiative cooling rate at these levels. Using a radiative-convective model with a physical parameterization for tropical convection, we find that only a significant increase of relative humidity in the middle and upper troposphere (20–40% greater than at present) can lead to the needed increase of lapse rate in the low troposphere. It is noted that the profiles observed in CLIMAP have much more convective available potential energy (CAPE) than present profiles, which can lead to enhanced production of hydrometeors in the middle and the upper troposphere whose reevaporation can moisturize the free troposphere. It is shown that the moisture change leads to an extremely strong negative feedback which greatly reduces the response to a doubling of CO_2. Finally, we discuss the apparent contradictions between the present results and existing studies suggesting a positive water vapor feedback.

1 Introduction

Given present concerns about climate, the historical climate record assumes special importance. In this connection, our ability to understand past climates is an important test of our understanding of climate in general. Moreover, proxy records from the distant past enable us to consider markedly different climate regimes from the present, while instrumental records only cover periods differing modestly from the present. Notable among paleoclimatic records is the CLIMAP reconstruction of the Earth's climate 18 000 years ago (ybp) (CLIMAP, 1976) during the last major glaciation. A remarkable finding of this extensive study was that the average tropical sea surface temperature was only about 1 °C colder than at present. Equally remarkable was the finding that the snow line on high mountains descended about 1 km – even on equatorial peaks (Broecker and Denton, 1989; Rind and Peteet, 1985). This represents a cooling of about 5 °C at about 5 km (Rind, 1990; Rind and Peteet, 1985), and hence a 20% increase in lapse rate below this level.

The 1 km descent of the mountain snowline was found over widely dispersed geographic regions and the lowering on the wet side of the mountains was about the same as on the adjacent dry side (Broecker and Denton, 1989). This suggests that the descent of the mountain snowline is unlikely to have been caused by changes in the local circulations. More importantly, in the equatorial tropics, the Rossby radius of deformation is extremely large, and temperatures above the trade wind boundary layer are virtually horizontally uniform. The dynamics appropriate to the equatorial tropics simply does not permit large gradients to develop above the trade wind boundary layer (especially over the large averaging periods pertinent to the paleoclimatic record). As a result, the temperature decrease observed on mountains is likely to be characteristic of the tropics. We do not yet have a detailed analysis of the instrumental record for the response of the static stability to the increases in globally averaged surface temperature during the late 1970's, although Angell (1988) does suggest some enhanced warming for the 850-300 mb layer. As concerns the paleoclimatic record, all that can be said is that the CLIMAP program involved a large part of the paleoclimatic community, and the data was a carefully and critically assessed as is currently possible.

[1] *Current address*: Geophysical Fluid Dynamics Laboratory, Princeton University, Princeton, NJ 08542, USA
Correspondence to R. S. Lindzen

Uncertainties on the order of 1 °C are possible, but confidence in the data is generally high. Our approach in this paper is to take the data seriously, and see what it implies. As always, it is possible that the data, itself, is wrong, but it would seem unreasonable to start with that supposition. This paper shows that the altered temperature profile of the last major glaciation required increased infrared opacity in the upper troposphere implying, in turn, increased humidity and, perhaps, cloudiness – at least in the tropics.

For reasons referred to in the preceding paragraph, we will concentrate on the tropics in this paper. The horizontal uniformity of the tropical temperatures allows one to average over the entire Hadley circulation, leaving advection only by the subsidence compensating the cumulus mass flux. Thus, following Schneider (1977), one can calculate the vertical thermodynamic structure of the tropical atmosphere on the basis of a one dimensional radiative-convective model. It should be noted immediately, that the above behavior of the lapse rate during the last glaciation is incompatible with simple notions of convective adjustments using either fixed lapse rates or moist adiabats. A moist adiabatic profile permits lapse rate change, but the change is too small to account for the lapse rate in the last glaciation. For a surface relative humidity of 80% (a good value for the tropics which is not sensitive to small temperature differences), a moist adiabatic profile gives about a 1.5° reduction of temperature at 5 km for a 1 K surface temperature decrease (say from 300 K to 299 K). It should be emphasized that both the notions of constants lapse rate and moist-adiabatic adjustment are empirical. Though the temperature profile within clouds is close to moist adiabatic (but not exactly) and probably remains so when climate changes, there is no a *priori* reason to believe this is the temperature of the mean field in which only a small percentage area is covered by clouds. The departure of the present tropospheric temperature from a moist-adiabat may be regarded as small, but the departure may change somewhat when the climate regime changes.

In this article, we consider a free troposphere containing deep convection coupled to a convective boundary layer. The parameterization for deep convection is based on Arakawa's mass flux formulation (Arakawa and Schubert, 1974), but the total net convective mass flux is determined through the moisture budget of the surface boundary layer (Lindzen, 1981, 1988). This closure has been previously used for climate studies (Lindzen *et al.*, 1982). A related parameterization has been used for forecasting (Geleyn *et al.*, 1982). The model is described in detail in Section 2. It is shown that this model successfully depicts the main characteristics of the observed vertical structure of the tropical tropospheric temperature.

The vertical structure of the tropical atmosphere is determined primarily by a balance between subsidence heating, $M_c \frac{\partial s}{\partial z}$ (where s = dry static energy, and M_c = cumulus mass flux), and radiative cooling. Since M_c, averaged over the tropics, is determined approximately by integrated evaporation divided by surface specific humidity – a quantity not expected to change significantly in a slightly cooler tropics, changes in $\frac{\partial s}{\partial z}$ must be balanced by reduced radiative cooling. Intuitively, this would appear to call for increased infrared opacity in the middle and upper troposphere. While one cannot completely reject the possibility that M_c might adjust to compensate for the reduction in $\frac{\partial s}{\partial z}$, this would call for a significant increase in M_c 18 000 ybp.

In order to address the above suggestion, we try, in Section 3, to simulate the vertical temperature structure during the last glaciation with various assumptions about the relative humidity profile in the middle and upper troposphere. It is shown that only a higher relative humidity than at present in the middle and upper troposphere gives a greater lapse rate in the low atmosphere than would a moist-adiabatic profile. The higher the relative humidity, the greater the lapse rate. Consistent with our original hypothesis, it is shown that the change in the distribution of radiative cooling rate accompanying a higher middle and upper tropospheric relative humidity is indeed characterized by a reduction in the low troposphere. In addition, there is enhanced cooling in the upper troposphere.

In Section 4 we examine the processes maintaining the tropical water vapor budget above the convective boundary layer, in order to see how cooler surface temperatures could lead to a higher relative humidity. This matter is discussed in greater detail in Sun and Lindzen (1992). In brief, we note that the troposphere above the trade wind boundary layer is moisturized largely by the re-evaporation of hydrometeors in the atmosphere (consistent with Gamache and Houze, 1983; Betts, 1990), and that the production of these hydrometeors (water and ice particles or drops) by cumulonimbus towers appears to increase with increasing CAPE (conditional available potential energy) (consistent with Szoke *et al.*, 1986; Rutledge *et al.*, 1992; Williams, *et al.*, 1992). We show that temperature profiles with the observed lapse rates do, in fact, have substantially greater CAPE than the present tropical atmosphere has.

In Section 5 we examine the implications of the inferred dependence of water vapor on surface temperature for climate sensitivity. We find that, in contrast with existing results for the water vapor feedback, the water vapor feedback is now strongly negative, virtually eliminating the warming due to a doubling of CO_2 – at least in the tropics. By comparing sensitivity studies from the present model with a model with a constant tropospheric lapse rate, we show the feedback from the lapse rate change is also considerable.

In Section 6 we discuss the apparent contradiction between the present result and existing results in the literature claiming to demonstrate a positive water vapor feedback. We note that model results are based on models which treat upper level water vapor in an unsatisfactory manner. We also note that existing efforts to use location and season as surrogates for climate to infer the water vapor feedback from data suffer from two defects: 1) insofar as air is subsiding over most of the globe, the upper level water vapor is not directly related to surface temperature, and 2) within the tropics, for a given climate, CAPE is largely related to local temperature in the boundary layer where horizontal heterogeneity is permitted, whereas under conditions of global climate change, tropics-wide CAPE changes also depend on the horizontally homogeneous temperature distribution above the boundary layer.

We finally address the question of what might constitute appropriate data for determining the relation of upper level water vapor to global climate.

2 The one-dimensional radiative-convective model

We adopt the simple average approach to the tropical circulation, first proposed by Schneider (1977) (see also Sarachik, 1978). This approach is schematically illustrated in Fig. 1. The vertical structure is characterized by a convective boundary layer, a free troposphere and the atmosphere above.

In the troposphere above the convective boundary layer, the convective heating, averaged over the tropics (i.e., the Hadley domain), results from cloud-induced subsidence heating. The thermal energy equation for an averaged radiative-convective equilibrium is

$$M_c \frac{ds}{dz} - R_e = 0 \tag{1}$$

for $z_b < z < z_t$. M_c is the cloud-induced subsidence, s is the dry enthalpy (static energy), R_e is the radiative cooling rate, z_b is the height of the top of the boundary layer and z_t is the height at which the convective heating ends – essentially the tropopause.

Since convective heating ends at the tropopause level, in a radiative-convective equilibrium, the net radiative flux across the top of the tropopause and the top of the atmosphere are both zero. The integral form of the thermal energy equation for the free troposphere can be written as

$$\int_{z_b}^{z_t} M_c \frac{ds}{dz}\, dz = S{\downarrow}_b - F{\uparrow}_b \,. \tag{2}$$

$S{\downarrow}_b$ and $F{\uparrow}_b$ are the net solar flux and the net infrared flux across the top of the convective boundary layer.

Within the convective boundary layer, the dry enthalpy is assumed to be well mixed. A finite jump of temperature across the top of the boundary layer is allowed and it is related to the surface water vapor flux

$$M_c \Delta s = \alpha L E_s \,. \tag{3}$$

L is the latent heat, E_s is the surface water vapor flux and α is the entrainment ratio (Sarachik, 1985).

The integral form of the heat budget of the boundary layer is

$$M_c \Delta s + F_s = (S{\downarrow}_0 - F{\uparrow}_0) - (S{\downarrow}_b - F{\uparrow}_b) \,. \tag{4}$$

F_s is the sensible heat flux from the surface; $S{\downarrow}_0$ and $F{\uparrow}_0$ are the net solar flux and infrared flux across the surface. The surface heat budget is

$$S{\downarrow}_0 - F{\uparrow}_0 = L E_s + F_s \,. \tag{5}$$

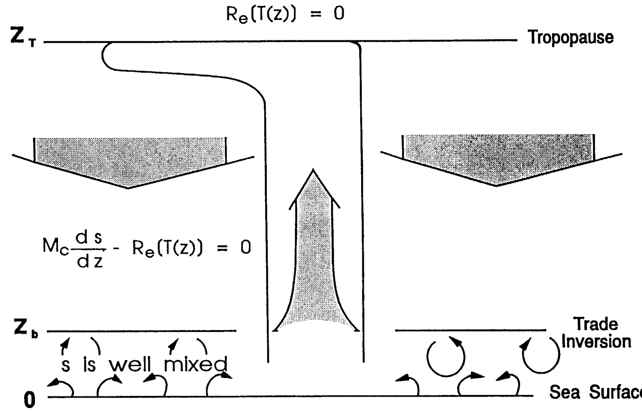

Fig. 1. A schematic illustration of the model structure. z is the height, T is the temperature, R_e is the net radiative cooling, s is the dry enthalpy (static energy), M_c is the cloud induced subsidence, z_t is the height of tropopause and z_b is the height of the convective boundary layer

From Eqs. (2), (4) and (5) we obtain

$$\int_{z_b}^{z_t} M_c \frac{ds}{dz}\, dz + M_c \Delta s = L\ E_s \,. \tag{6}$$

Following Lindzen *et al.* (1982), the tropopause height, z_t is given by the neutral buoyancy level:

$$s(0) + r h_s L q^*(0) = s(z_t) \,, \tag{7}$$

where $s(0)$ is the dry enthalpy of surface air, $s(z_t)$ the dry enthalpy at the tropopause level, $q^*(0)$ the saturation water vapor mixing ratio of surface air, and $r h_s$ the relative humidity of surface air participating in deep convection. The last quantity may not be the same as, and is likely larger than, the mean surface relative humidity. Calculations showed that when $r h_s$ is assumed to be 1, there is a better correlation between the tropopause potential temperature and the left side of Eq. (7) (Chimonas and Rossi, 1987). Here we assume $r h_s$ equals 1 for simplicity. Note that Eq. (7) is only approximate. Such factors as the detrained humidity have been neglected since their contribution to buoyancy is negligible. In fact, one can not go too far in relating the cloud environment to the mean environment, since there are important differences; for example, the trade inversion is unlikely to be seen by the deep clouds.

Budgets studies show that above the convective boundary layer, M_c is roughly constant with height up to 300 mb and then sharply decreases to zero (Ogura and Cho, 1973; Cheng *et al.*, 1980). For simplicity, we take M_c to be constant with height in the free troposphere. From Eqs. (6) and (7) we have the following equation for M_c,

$$M_c = \frac{E_s}{r h_s q^*(0)} \,. \tag{8}$$

The air-sea temperature difference is assumed to be 1 K, close to the climatological mean value. There are good reasons to believe this difference cannot be large since the heat capacity of air is very small. Note that no explicit formula for surface flux is needed after the air-sea temperature difference is specified. When the aerodynamic formula is employed, the air-sea temperature difference can be calculated. However, the drag coefficient may have to be adjusted to give a reasonable air-sea temperature difference.

The radiative code used, designed by M. D. Chou, is a band model including water vapor, CO_2, O_3, CH_4 and N_2O. No aerosol is taken into account, and at this point, we have little basis for realistically considering its variability with climate. The performance of this code in calculating the cooling and heating rates has been compared with other radiation codes frequently used in climate sensitivity studies, and also with the most sophisticated line by line calculation (Chou, 1986; Chou *et al.*, 1991). In dealing with water vapor, this code is found to give excellent agreement with the results obtained by the line by line calculations.

Figure 2 presents two equilibrium solutions[1] with fixed specific humidity and fixed relative humidity. CO_2 and other chemical species are specified at their present values. In the calculation, no clouds are explicitly included. The solar insolation is adjusted to give the surface air temperature, 300 K; this implicitly accounts for effects of clouds on albedo and heat flux out of the tropics.

The model profile is very close to the observed as far as gross features are concerned. The tropopause level determined by the neutral buoyancy assumption is 180 mb and 160 mb respectively for the two states (160 mb for fixed relative humidity and 180 mb for fixed specific humidity). This level should be interpreted as the mean level at which deep clouds detrain when compared to observations (the observed mean detrainment level of deep clouds is about 200 mb). The observed tropopause level is somewhat higher because it is determined by the tallest clouds which originate at the hottest spots in the tropics (note that dynamics, by smoothing horizontal temperature gradients, establishes the tropopause throughout the tropics).

The temperature around the tropopause level is slightly higher than observed, primarily because the tropopause height is underestimated. Additionally, the profile of humidity used in the calculations in the upper troposphere may be inaccurate. The tropopause height and temperature are found to depend on the humidity in the upper troposphere and stratosphere, but there are no adequate observations of water vapor there. Further, assuming M_c is constant with height up to the tropopause is not very realistic. Finally, the effect of upper level clouds on the infrared radiative cooling rate is not included in the radiative code. By adjusting the water vapor profile in the upper troposphere and stratosphere or by choosing a profile for M_c in the upper troposphere which decreases to zero over a finite depth (say from 300 mb to the tropopause), one can indeed get at better fit with observations. However, precise tuning of what are, for present purposes, unessential details is not our purpose. What is important here is that the model is able to grasp the main features of the tropical troposphere in an energetically consistent (potentially unstable) way. It also gives a simple interactive picture of how the tropical tropospheric temperature is maintained. Figure 3 shows the vertical structure of the saturation equivalent potential temperature for the two equilibrium states; it shows that

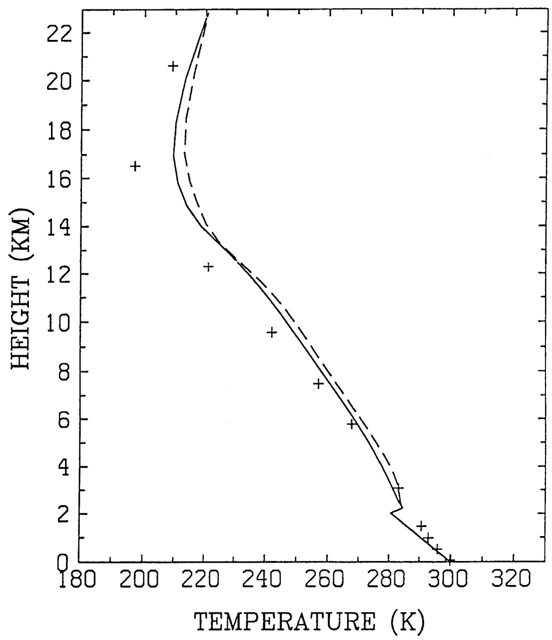

Fig. 2. Equilibrium temperature. *Solid line:* fixed relative humidity. *Dashed line:* fixed specific humidity. The relative humidity is specified as $rh(p)=0.8\,\frac{(p/1000)-0.02}{(1-0.02)}$; p is the pressure. The minimum mixing ratio of water vapor is set to be 0.03 g/Kg. The specific humidity profile is from the standard tropical troposphere (McClatchey *et al.*, 1972). +'s indicate the observed temperature averaged over the domain of the Hadley circulation in January. The data are from Oort (1983)

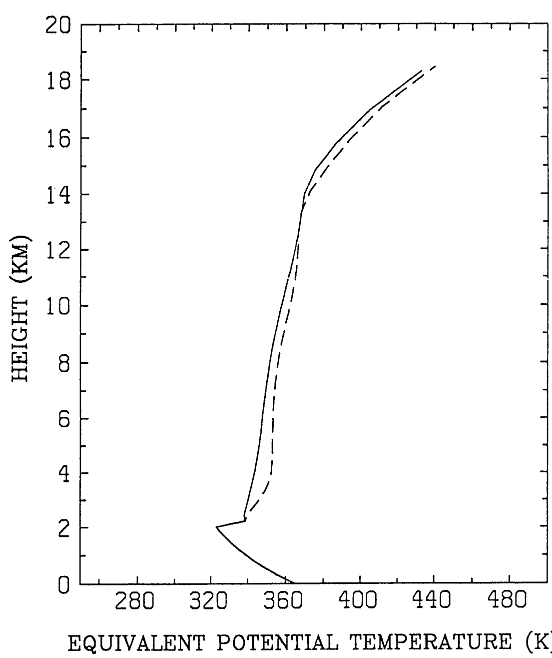

Fig. 3. Equilibrium saturation equivalent potential temperature, θ_e^*. *Solid line:* fixed relative humidity; *dashed line:* fixed specific humidity

[1] All our calculations of temperature have used a vertical resolution of 20 mb up to the level of 20 mb. Above 20 mb, there are only three additional levels: 10 mb, 1 mb, and 0.1 mb

the equilibrium profile is potentially unstable. Above the boundary layer, the saturated equivalent potential temperature is almost constant with height, which indicates the temperature profile is close to moist-adiabatic. The minimum of θ_e^* occurs at the top of the convective boundary layer. The observed profile of θ_e^* is smoother than the model profile. In reality, the height of the boundary layer oscillates and shallow clouds are not strictly restricted below the climatological mean height for the convective boundary layer. Nevertheless, it is the presence of the boundary layer that maintains a potentially buoyant tropical troposphere. The model is designed to reflect this as simply as possible.

It should be emphasized that for present calculations, the calculated profile is close to the moist-adiabat as is evident in Figs. 2 and 3. This is hardly surprising given that the surface air temperature, $T(0)$, and the tropopause temperature lie on the same moist-adiabat (Eq. (7)). However, local (in height) deviations are allowed in our model. The deviations are dependent on the radiative cooling rate and the magnitude of M_c. This feature is crucial for dealing with implications of the depressed snowline during the last glaciation.

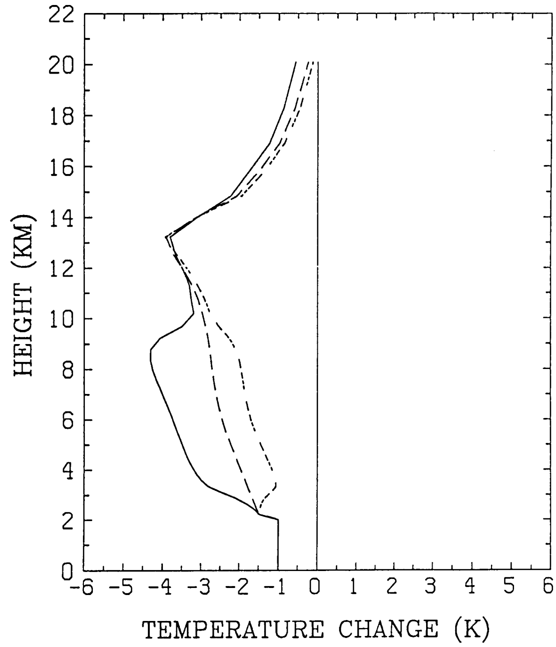

Fig. 4. Temperature change as a function of height when the surface temperature is decreased by 1 K. *Long-dashed line:* Relative humidity is fixed and is assumed not to change when the surface temperature is decreased by 1 K. *Solid line:* Relative humidity between 300 mb and 700 mb is assumed to increase by 0.2 when the surface temperature is decreased by 1 K. *Short-dashed line:* Relative humidity between 300 mb and 700 mb is assumed to be 80% of the present value

3 Numerical experiments

The model presented in the preceding section shows how the vertical structure of tropospheric temperature is calculably related to the distribution of radiative cooling rate. This allows us to infer, from the observed temperature change, the tropical water vapor distribution during the last glaciation.

In many past climate sensitivity studies, relative humidity throughout the troposphere is assumed to be independent of climate change. As pointed out by Lindzen (1990) and Betts (1990), little is concretely known about the maintenance of the moisture budget in the free troposphere (i.e., the troposphere above the boundary layer) and its sensitivity to radiative perturbation. In attempting to simulate the tropospheric temperature profile during the last glaciation, we adjust the relative humidity in the free troposphere in order to obtain a vertical temperature structure which resembles the one indicated by the mountain snowline record. Since observations show that the relative humidity within the convective boundary layer (up to about 700 mb) is relatively insensitive to surface temperature change, we fix the relative humidity below 700 mb in the following calculations. It should be emphasized that the moisture budget below 700 mb is largely determined by turbulence driven by surface fluxes (Sun, 1992). Model results indicate that relative humidity in the boundary layer does not depend significantly on temperature changes of a few degrees (Betts and Ridgway, 1989), and this is consistent with observations (Oort, 1973). On the other hand, the dependence of the relative humidity in the free troposphere cannot be related to the local sea surface temperature. Over much of tropical ocean, the influence of local sea surface temperature is restricted to below the trade inversion; away from active cumulus towers (which occupy only about 0.1% of the tropical area), air above the trade inversion is subsiding, and the temperature and moisture fields are largely related to the temperature and moisture fields over the hottest points where deep convection occurs. We will return to this point in our final section where we discuss claims of a positive feedback due to water vapor.

As already noted, we use "solar flux" as a surrogate for the effects of cloud albedo and heat export from the tropics. We adjust this flux to produce equilibrium profiles with both the present surface temperature and a surface temperature 1 K colder than at present. Different relative humidity profiles in the middle and upper troposphere are assumed to accompany the colder climate. Figure 4 shows three cases of the temperature departure from the present climate as a function of height. The long-dashed line is the case in which the relative humidity is assumed to be the same as the present. The short-dashed line represents the case in which the relative humidity between 300 mb and 700 mb is only 80% of the present value. The solid line represents a case in which the relative humidity between 300 mb and 700 mb is 20% larger than the present profile (i.e. present relative humidity plus 20%)[2].

From Fig. 4, we see that when the relative humidity is assumed not to have changed when the surface temperature is decreased by 1 K, the temperature reduction in the middle troposphere is about 2 K – basically the behavior

[2] 160 mb, the cloud detrainment level can be used instead of 300 mb. There is no substantial difference in the results. On physical grounds, we expect little change of relative humidity at the detrainment level. See Section 4 for further discussion

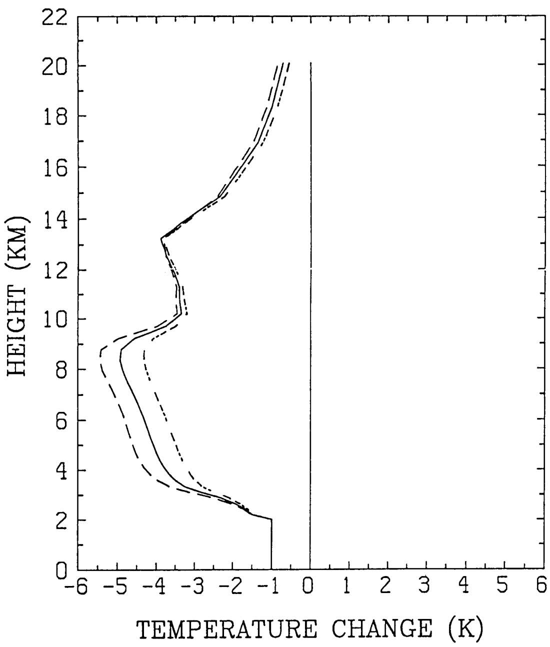

Fig. 5. Temperature change as a function of height when the surface temperature is decreased by 1 K. Relative humidity between 300 mb and 700 mb is assumed to increase to a specified value when the surface temperature is decreased. *Short dashed line:* the relative humidity is assumed to increase by 0.2. *Solid line:* the relative humidity is assumed to increase by 0.3. *Long dashed line:* the relative humidity is assumed to increase by 0.4

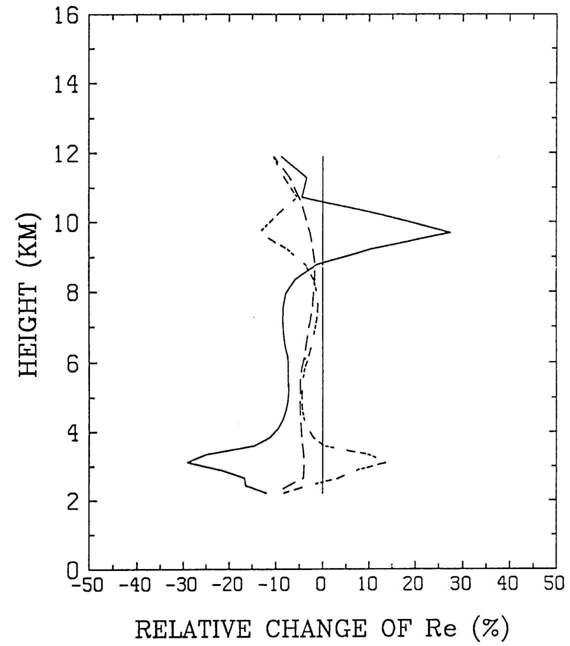

Fig. 6. Relative change of tropospheric radiative cooling as a function of height when the surface temperature is decreased by 1 K. Different assumptions about the change of the relative humidity between 700 mb and 300 mb are made. *Solid line:* the relative humidity is assumed to increase by 0.2 *Long dashed line:* the relative humidity is assumed not to change. *Short dashed line:* the relative humidity is assumed to be 80% of the present value

is the same as that obtained by assuming a moist-adiabat. When a smaller relative humidity in the free troposphere is assumed to accompany the colder climate, the reduction of temperature in the middle troposphere is actually less than the surface temperature decrease. When the relative humidity in the free troposphere is assumed to be 20% higher than at present, the reduction at the middle troposphere is about 3.5 K, which is about the 70% of the reduction indicated by the mountain snowline record.

From the three cases, it is clear that only an increase of relative humidity in the free troposphere can lead to a greater reduction of temperature in the middle troposphere than what a moist-adiabatic profile predicts. It is further found that the steepening of the lapse rate in the free troposphere is proportional to the increase of relative humidity in the free troposphere, though not in a linear fashion. Figure 5 shows the departure of temperature from the present as a function of height when even higher relative humidity is assumed to accompany the colder climate. To obtain a 5 K temperature reduction at the mid-troposphere, the increase of relative humidity needs to be about 40%. There may, however, be reasons for this estimate being excessive, which will be discussed shortly.

Physically, the changing lapse rates are associated with changes in radiative cooling below about 6 km (viz. Eq. (1)). Increasing humidity decreases local cooling rates by increasing the infrared opacity above, and decreasing cooling rates are associated with decreasing ds/dz or,

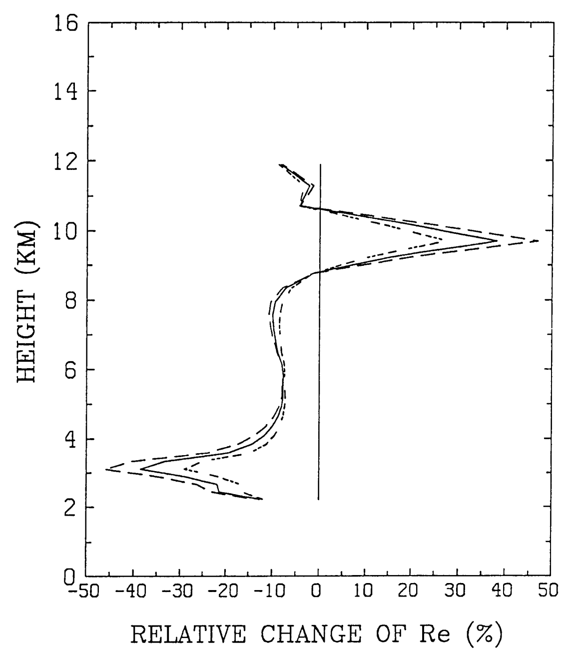

Fig. 7. Relative change of net tropospheric radiative cooling as a function of height when the surface temperature is decreased by 1 K. Relative humidity between 300 mb and 700 mb is assumed to increase to a specified value. *Short-dashed line:* the relative humidity is assumed to increase by 0.2. *Solid line:* the relative humidly is assumed to increase by 0.3. *Long-dashed line:* the relative humidity is assumed to increase by 0.4

equivalently, increasing lapse rate. Figures 6 and 7 show the change in the distribution of radiative cooling rate for various choices of humidity distribution. M_c and the tropopause height determined by the neutral buoyancy level are both found to be almost the same in both the present climate and the colder climate for all cases. When there is more water vapor in the middle and upper troposphere, the radiative cooling rate change is characterized by an increase in the upper troposphere (due to enhanced emissivity) and a reduction in the lower troposphere (due to increased opacity above). The temperature reduction maximizes in the middle troposphere. These results are not sensitive to the assumption of fixed rh_s in the surface boundary layer (Sun, 1992).

Increased upper level cloud coverage can also contribute to the change in the distribution of radiative cooling rate. Indeed, considerable increase in the upper cloud amount, either in the form of increased thickness or wider horizontal coverage, is expected to accompany an increase of relative humidity in the free troposphere. It has been found that the free troposphere is largely moisturized by evaporation of hydrometeors falling from the upper level clouds (Sun and Lindzen, 1992). In this sense, the increase of relative humidity may be somewhat smaller than 40%. However, due to the uncertainties in the radiative properties of clouds, it is premature to quantify the effects with confidence. All that can be said at this point is that upper level cloud amount is not expected to increase without an increase of relative humidity in the free troposphere. Increased upper level cloudiness will also be associated with increased albedo. This may prove very important. As we have already noted, we have adjusted the solar constant in order to produce the observed tropical sea surface temperature. The necessary reduction was 4.6% when the relative humidity was taken to be 40% greater, and 3.1% when the relative humidity was taken to be 20% greater. The intimate coupling of cloud cover and humidity and the fact that they tend to influence climate in both complementary and opposite ways indicates the complexity of the situation.

Another reason that the actual increase of water vapor may be less than 0.4 is that a more moist atmosphere would facilitate the descending of the snowline (Rind, 1990). Though the controlling factor for the position is the temperature, relative humidity may also matter. Thus, once the relative humidity in the middle and upper troposphere is considerably higher than at present, the temperature reduction corresponding to 1 km descent may be less than 5 K.

The non-uniform temperature change leads to an interesting feature of the colder states: namely, the air participating in deep convection will experience enhanced buoyancy. In other words, the environmental profile has more convective available potential energy (CAPE). CAPE is simply the work done by buoyancy on a parcel of surface air rising in a cloud. This is proportional to the integral over the cloud of the difference between the equivalent potential temperature of cloud and ambient air. Our calculations of equivalent potential temperature follow Bolton (1980). Table 1 summarizes the CAPE of the different profiles discussed above. The steeper the lapse rate, the larger the CAPE. The increase of CAPE during the last glaciation is 40% to 100%, and suggests why the relative humidity in a colder climate could have been higher than at present. We will explore this point in the following section.

Table 1. Conditional Available Potential Energy (CAPE) associated with various assumed distributions of relative humidity

T_s (K)	ΔR	CAPE (joules) $rh_s=0.8$	CAPE (joules) $rh_s=1.0$
300	0	670.5	2200.3
299	+0.2	1047.2	2610.9
299	+0.3	1156.8	2749.7
299	+0.4	1255.9	2841.0

4 A one-dimensional model for the tropospheric relative humidity

Consistent with the one-dimensional, meridionally-averaged model for the temperature structure, the budget of tropospheric water vapor above the convective boundary layer can be written as

$$M_c \frac{dq}{dz} + \varrho\, E_r = 0 \tag{9}$$

for an equilibrium state. q is the water vapor mixing ratio. $M_c \frac{dq}{dz}$ is the downward moisture advection due to deep-cloud induced subsidence. ϱ is the air density and $\varrho\, E_r$ is the moisturization from deep convection. The moisturization from deep convection is largely from evaporation of hydrometeors (Sun and Lindzen, 1992). Hydrometeors here refers to ice or liquid water existing outside of saturated cumulus updrafts. That the source of water vapor above the boundary layer is the evaporation of water droplets and ice crystals falling outside of cumulus towers, but originating in the ice and water detrained from these clouds, is increasingly recognized (Gamache and Houze, 1983; Betts, 1990). This result is readily understood by the following considerations. First, it is virtually impossible to account for mid-tropospheric humidity simply through the subsidence of saturated air detrained from the tops of deep clouds. Due to adiabatic heating, the relative humidity of such air will decrease to below 5% by the time the air has descended a few kilometers – contrary to observations. Second, if one attempts to transport water vapor upward by large-scale turbulent or laminar flow one encounters the fact that such air will, within a kilometer or so of its origin, saturate because of adiabatic cooling, and that such a process cannot transport water vapor deep into the troposphere without producing an extensive and deep saturated layer – something also not observed on earth. Moisturization via evaporation of falling condensate is free of these problems.

The moisturization from evaporation of hydrometeors may be parameterized as $E_r = \alpha(q^* - q)$, in which q^* is the saturation water vapor mixing ratio and α is a coefficient which is related to hydrometeor water content (total mass of hydrometeors per unit volume) and the spectrum of

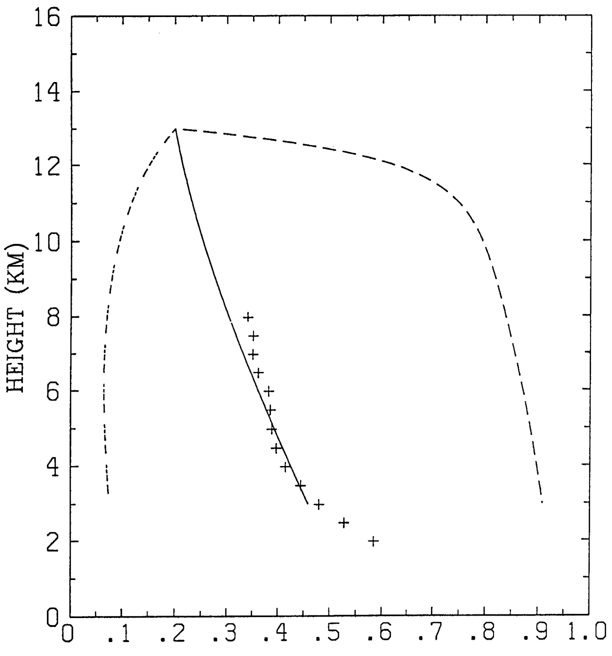

Fig. 8. Vertical distribution of relative humidity with different choices of h_2. The cloud top is at 13 km and relative humidity there is assumed to be 1. h_2 for the three curves is: 0.20 km (*long dashed line*), 2.0 km (*solid line*) and 20.0 km (*short dashed line*). +'s correspond to the data averaged over the domain of Hadley circulation in January (Oort, 1983). h_3 is chosen as 3 km

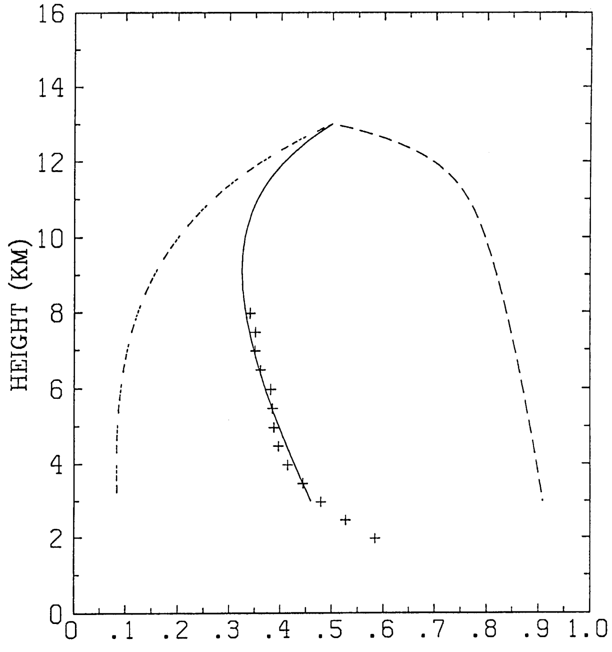

Fig. 9. Same as Fig. 8, but the relative humidity at the cloud top level is assumed to be 0.5

hydrometeors[3]. The moisture budget equation for the free troposphere can be rewritten as

$$M_c \frac{dq}{dz} + \alpha \varrho (q^* - q) = 0 . \tag{10}$$

The above equation for the specific humidity can be rewritten as an equation for the relative humidity, $R = q/q^*$,

$$\frac{dR}{dz} - \frac{R}{h_1} + \frac{1}{h_2} \exp(-z/h_0) = 0 . \tag{11}$$

h_1 and h_2 are two parameters with dimensions of height, defined as:

$$\frac{1}{h_2} = \frac{\alpha \varrho_0}{M_c} \tag{12}$$

$$\frac{1}{h_1} = \frac{1}{h_3} + \frac{1}{h_2} \exp(-z/h_0) \tag{13}$$

$$\frac{1}{h_3} = -\frac{d \ln q^*}{dz} . \tag{14}$$

h_0 is the density scale-height of the air and ϱ_0 is the density of surface air.

To a good degree of approximation, q^* decreases exponentially with height; therefore, h_3 is a constant. Now we see that R is not explicitly dependent on temperature. However, this does not mean the relative humidity in a warmer or colder climate should be the same. The parameter h_2 and thus h_1 may be very different in cold and warm climates because they are related to microphysical processes in deep clouds. By definition, we know that h_2 is a parameter measuring the relative strength of the two physical processes: the removal of water vapor by cloud induced subsidence, and the deposition of water vapor through the moistening processes.

Figures 8, 9 and 10 show the vertical distribution of the relative humidity with different choices of h_2 and different choices of relative humidity at the detrainment level[4]. If there is no lateral transport and no exchange of water vapor with the stratosphere, the relative humidity at the detrainment level should be 1. In reality, both lateral transport and exchange with the stratosphere occur; the relative humidity at the detrainment level can thus be less than 1. In the absence of adequate observations of water vapor in the upper-troposphere, there is little empirical basis for any choice. We therefore consider various values of relative humidity at the detrainment level. Fortunately, we find that this uncertainty does not affect the present analysis. In Fig. 8, the relative humidity at the detrainment level is fixed at 1, while in Figs. 9 and 10, the relative humidity at the detrainment level is fixed at 0.5 and 0.2 respectively. Three curves are presented in each figure, corresponding to different choices of h_2. There are two remarkable features in these figures. First, away from the vicinity of the cloud top, the vertical structure of the relative humidity is not sensitive to the relative humidity

[3] α has been used earlier (Equation 3) to represent the entrainment ratio at the top of the tradewind boundary layer. There should be no confusion since the usage is invariably clear from the context

[4] All calculations of humidity used a vertical resolution of 100 meters

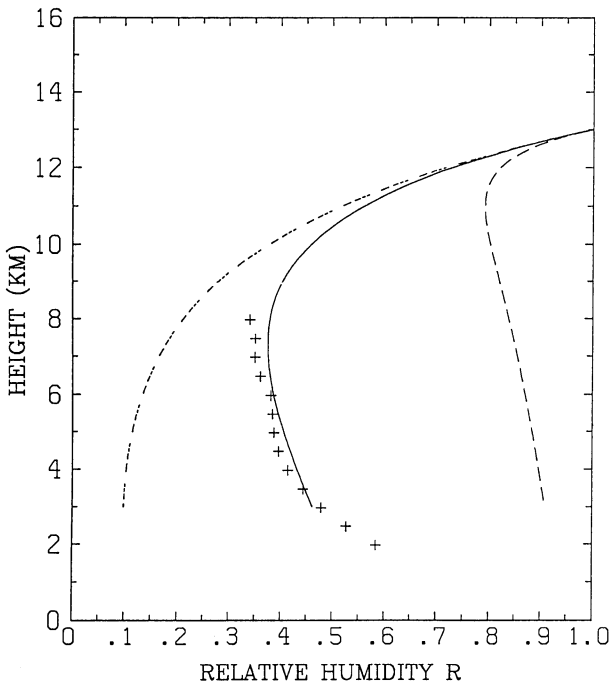

Fig. 10. Same as Fig. 8, but the relative humidity at the cloud top level is assumed to be 0.2

at the level of the cloud top. Second, between 700 mb and the height (roughly 400 mb) below which we have data, the vertical structure corresponding to the solid curve closely resembles the observed vertical distribution of relative humidity.

In the context of this model, we see that the relative humidity is indeed strongly controlled by h_2. Significantly decreasing h_2 leads to increases in relative humidity comparable to the magnitude inferred from the mountain snowline record.

As shown in the radiative-convective equilibrium calculations, M_c changes little (viz. Eq. (8)). Therefore, the change of h_2 has to arise from the change of α, which is related to the hydrometeor water content and also the spectrum of hydrometeors. To be more specific, we consider an idealized case here. For hydrometeors with a uniform spectrum and a spherical shape, α may be written as

$$\alpha = 3 D C_v \frac{\varrho_r}{\varrho_L} r^{-2}, \qquad (15)$$

where D is the diffusion coefficient for water vapor in the air, C_v is the ventilation coefficient, ϱ_r is the hydrometeor water content, ϱ_L is the density of hydrometeors and r is the radius of hydrometeors (Betts and Silva Dias, 1979). It follows that h_2 is given by

$$h_2 = \frac{\varrho_L}{\varrho_0} \frac{M_c r^2}{3 D C_v \varrho_r}, \qquad (16)$$

so that a smaller h_2 requires a larger amount of smaller hydrometeors.

Accompanying the deepening of the lapse rate, there was much more CAPE in the last glaciation (about a doubling relative to the profile representing the present climate). Deep convection, associated with larger CAPE is observed to generate more hydrometeors in the middle troposphere and upper troposphere (Williams *et al.*, 1992). Sage II measurements also show that cirrus occur much more frequently over the tropical land than over the tropical ocean (Chiou *et al.*, 1990). It is known that deep convection over the land is associated with more CAPE than over the ocean. That the amount of hydrometeors present in the middle and upper troposphere increases with increasing CAPE is also expected on the basis of results from comprehensive numerical cloud models (M. K. Yau, personal communication, 1991). While the quantitative relationship between the amount of hydrometeors and the environmental CAPE has not yet been firmly established, the sign is not in serious doubt. That detrained water and ice should be proportional to CAPE is entirely reasonable. For small CAPE, cloud air will rise slowly, allowing warm rain processes to operate so that a large part of the water condensing within the cloud to fall out within the cloud in the form of rain. However, with larger CAPE, cloud updrafts will be larger, allowing less time for warm rain processes. Thus, the updrafts will carry more condensed water aloft, leading to more detrained water and ice (since much of the detrainment occurs above the freezing level).

Besides the change in the amount of hydrometeors, the spectrum of hydrometeors is also expected to change. When the environment has more CAPE, convective updrafts will be stronger. The hydrometeors carried into the middle and upper troposphere will have smaller sizes because they will have less time to grow before they reach the middle and upper troposphere.

5 Estimating the feedback factor of water vapor and lapse rate in doubling CO_2

We have, thus far, estimated the change in humidity in the tropical free troposphere accompanying a 1 K decrease in tropical surface temperature. Since our calculations are specific to the tropics, we cannot readily use these results to infer the sensitivity of global climate. However, our results do enable us to estimate how much the sensitivity of tropical surface temperature to a doubling of CO_2 is affected by the inferred changes in relative humidity.

Noting that a 1 K cooling of the tropical surface lead to an increase in relative humidity of as much as 0.4, we will assume that other tropically averaged surface temperature changes result in proportional changes in relative humidity. For reasons already discussed, a 40% change in tropical relative humidity for a 1 K change in average tropical surface temperature is probably an upper limit. Thus, in calculating the response of tropical surface temperature to a doubling of CO_2, we consider a range of values for the expected increase of relative humidity per 1 K decrease of surface temperature (ΔR_{-1}). The formula used is simply

$$\Delta R\,(300\,\text{mb} - 700\,\text{mb}) = -\Delta R_{-1} \times (T_s - 300\,\text{K}). \qquad (17)$$

Equation (17) tells us that when a doubling of CO_2 in our radiative code leads to a surface warming, the resulting increase in surface temperature will lead to a reduction of upper level relative humidity. Of course, a more appropriate approach would involve using something like Eq. (11) to explicitly calculate relative humidity. This would, however, require fairly sophisticated cloud modeling. Similarly, it will eventually be necessary to consider associated changes in albedo. Although we are engaged in such efforts, they go beyond the scope of the present paper. For the moment, Eq. (17) can be viewed as a linearization. To be sure, if ($T_s - 300$ K) proves large, this linearization would break down. However, as we see in the results summarized in Table 2, ($T_s - 300$ K) remains small for a doubling of CO_2. We see that for fixed specific humidity, we would expect an increase in tropical surface temperature of 0.66 K. However, for a 40% increase in R per 1 K decrease of average surface temperature the response decreases to 0.09 K, and even for a 10% increase of R per 1 K decrease of surface temperature, the response is still only 0.23 K. These values may be contrasted with results from models wherein a doubling of CO_2 produces a 4 K increase in global mean temperature. In these models, tropical temperatures increase by 2 K or more.

In Table 2 we also show feedback factors, f, defined by the relation

$$\text{response} = \frac{\text{response for fixed specific humidity}}{1-f} \tag{18}$$

These feedback factors, while germane to the tropical half of the earth, may not be applicable to the whole globe. For example, during the last glaciation, global mean temperatures were several degrees lower than at present even though tropical temperatures were only 1 K lower. On the other hand, glacial climate changes were almost certainly not associated with changes in the global radiative forcing, but rather with changes in the meridional heat flux and regional inhomogeneities in radiative forcing. In such a case, the change in global mean temperature is likely to be a peripheral consequence of the increase in equator-to-pole temperature gradient, and the relevance of simple feedback factors is dubious. However, for sensitivity to a globally homogeneous forcing such as doubling of CO_2, these factors could very well have a more general relevance, and the numbers in Table 2 may be indicative of global as well as tropical feedbacks.

It should be further noted that associated with a change in the relative humidity profile, there is a change in the lapse rate. As emphasized by Cess *et al.* (1990) and Cess (1991), the feedback from the lapse rate is potentially as important as the water vapor feedback, though the two always go together. To isolate the role of lapse rate change from the change of relative humidity in the free troposphere, we employ a radiative-convective model with constant lapse rate adjustment to investigate the role of water vapor feedback in double CO_2 experiments. The results are summarized in Table 3.

By comparing Tables 2 and 3, we see that the feedback factor from the lapse rate change is as large as that from the change of water vapor, and that both add. However, it should be emphasized that the two are dynamically coupled, though their roles in affecting radiative transfer in the atmosphere are physically different. Indeed,

Table 2. Tropical surface temperature change, ΔT, for a doubling of CO_2 and associated feedback factor, f, for different assumed changes in relative humidity (between 300 mb and 700 mb) for a 1 K decrease in sea surface temperature (ΔR_{-1}). Lapse rate permitted to vary

ΔR_{-1} (%)	ΔT (K)	f
Fixed q	0.66	0
10	0.23	−1.9
20	0.15	−3.5
30	0.12	−4.6
40	0.09	−6.4

Table 3. Tropical surface temperature change, ΔT, for a doubling of CO_2 and associated feedback factor, f, for different assumed changes in relative humidity (between 300 mb and 700 mb) for a 1 K decrease in sea surface temperature (ΔR_{-1}). Lapse rate held constant at 6.5 K/km

ΔR_{-1} (%)	ΔT (K)	f
Fixed q	1.1	0
10	0.74	−0.5
20	0.44	−1.4
30	0.32	−2.4
40	0.24	−3.6

Lindzen *et al.*, (1982) found a similar effect, noting that it was related to the fact that the physical parameterization of cumulus heating lead to a higher deposition of heat than is implicit in simple convective adjustment. It may be argued that the diminished response in Table 2 for a fixed q (relative to the results in Table 3 for a fixed lapse rate) represents a negative lapse rate feedback. We prefer to consider this a base case from which feedbacks due to changing water vapor are calculated. Considering the lapse rate changes as feedbacks would further increase the negative feedback estimates.

6 Summary and discussion

We used a radiative-convective model to investigate what changes of water vapor above the convective boundary layer can lead to a picture consistent with the vertical structure of the tropical tropospheric temperature indicated by the mountain snowline record, i.e., a steeper lapse rate in the low atmosphere (or equivalently a reduction in static stability) requiring a concomitant reduction in radiative cooling rate. We found that only an increase of relative humidity in the middle and upper-troposphere during the last glacial period can explain this change in cooling rate. We also investigated what physical mechanism would lead to an increased relative humidity above the convective boundary layer in the context of our meridionally averaged one dimensional water vapor model. We noted that the profile during the last glaciation, indeed had more CAPE, which would have lead to the production of more and smaller hydrometeors in the middle and upper troposphere – both of these factors would lead to the enhanced moisturization of the free troposphere, consistent with the humidity changes required to produce the steeper lapse rate in the low troposphere which, in turn, produced the enhanced CAPE.

We also investigated the implication of the increase of relative humidity in the free troposphere for a decrease in surface temperature for the atmospheric response to a doubling of CO_2. We found that for the plausible values of relative humidity change inferred from the snowline record, the feedback of water vapor severely reduces the warming induced by the increase of CO_2. It is further noted that the large negative feedback factor is partially due to the lapse rate change associated with the change of water vapor distribution.

The nature of water vapor feedback suggested by this study is contrary to what GCMs have predicted. The problems of GCMs in dealing with convective transport of water vapor was discussed in Sun and Lindzen (1992). In brief, the parameterization schemes used in GCMs for calculating convective transport were not originally designed to address the issue of water vapor and have not been significantly improved in this regard. In particular, they lack adequate treatments of microphysical processes in precipitation formation and dissipation (see also Emanuel, 1991). In addition, there are numerical problems (Rasch and Williamson, 1990).

There are some observational studies which have claimed a positive dependence between sea surface temperature and relative humidity in the free troposphere (Rind *et al.*, 1991; Raval and Ramanathan, 1989). These results were obtained by comparing observed values of relative humidity or other related physical quantities over different places in the present tropics. In order for such observations to be interpreted as climate feedbacks, it is *essential* that upper level humidity be directly related to surface temperature. However, above the convective boundary layer, air is *subsiding* and the physical properties of air including water vapor and temperature are strongly correlated over large distances through dynamic transport. Thus location (or season) cannot be used as surrogates for global climate change.

Further, the properties of the vertical profiles which determine CAPE locally differ from those which determine changes of CAPE for a differing climate regime. Locally, within a climate regime, CAPE depends primarily on surface temperature (which varies horizontally to a much greater extent than does tropospheric temperature above the boundary layer). On the other hand, changes in CAPE between climate regimes depends also on the midtropospheric temperature changes described in the present paper. Moreover, since such changes apply to the whole tropics, they affect the overall and not just the local CAPE.

Alternate attempts to explain the mountain snowline record exist (Betts, 1991). These generally assume a moist adiabatic temperature profile, require major changes in quantities like the surface wind, and, most important, require that the decrease in tropical SST 18,000 ybp was 2 °C or more. Whether changes in equatorial temperatures this large are possible has been contested (Fairbridge, 1991), but the possibility remains. Caution is also required in inferring negative feedbacks from the paleoclimate record, given that we still are unable to satisfactorily account for the ice ages themselves. Undoubtedly, changes in the heat flux from the tropics to higher latitudes are involved. However, other factors involving currently unknown matters like aerosol changes may also play a role.

Finally, we briefly discuss observations which might test the applicability and correctness of the present results. It should be clear that local correlations of humidity and temperature do not properly indicate a climate feedback. However, comparisons of globally (or tropically) averaged distributions of temperature and specific humidity should. In this connection, it might be useful to consider the period in the late 70's and early 80's when the globally averaged temperature increased by about 0.25 °C (Houghton *et al.*, 1990). If the results of the present paper are applicable to the issue of global climate change, then one might expect measurable reductions in globally averaged specific humidity to accompany the observed warming. In addition, one should see enhanced warming (compared to what might be expected from the moist adiabat) at 500 mb relative to the surface – at least in the tropics. It is interesting in this regard that Oort (personal communication, 1992) is currently conducting exactly such a study using conventional radiosonde data. The importance of the water vapor feedback to estimates of global change certainly warrants the most intensive study of these issues.

Acknowledgements. This research was supported by the National Science Foundation under Grant ATM-8520354, and by the National Aeronautics and Space Administration under Grant NAGW-525. Helpful discussions with E. R. Williams and M. K. Yau are gratefully acknowledged.

References

Angell, J. K., Variations and trends in tropospheric and stratospheric global temperatures, 1958–87. *J. Clim.*, **1,** 1296–1313, 1988.

Arakawa, A. and W. H. Schubert, Interaction of a cumulus cloud ensemble with the large scale environment, Part I. *J. Atmos. Sci.*, **31,** 674–701, 1974.

Betts, A. K., Greenhouse warming and the tropical water vapor budget. *Bull. Amer. Meteor. Soc.*, **71,** 1465–1467, 1990.

Betts, A. K., Tropical boundary layer equilibrium in the last ice-age. *J. Geophys. Res.*, **97,** 2529–2534, 1991.

Betts, A. K. and M. F. Silva Dias, Unsaturated downdraft thermodynamics in cumulonimbus. *J. Atmos. Sci.*, **36,** 1061–1071, 1979.

Bolton, D., The computation of equivalent potential temperature. *Mon. Weath. Rev.*, **108,** 1046–1053, 1980.

Broecker, W. S. and G. H. Denton, The role of ocean-atmosphere reorganization in glacial cycles. *Geochim. Cosmochim. Acta*, **53,** 2465–2501, 1989.

Cess, R. D., Positive about water feedback. *Nature*, **349,** 462, 1991.

Cess, R. D., G. L. Potter, J. P. Blanchet, G. J. Boer, A. D. Del Genio, M. Déqué, V. Dymnikov, V. Galin, W. L. Gates, S. J. Gahn, J. T. Kiehl, A. A. Lacis, H. Le Treut, Z.-X. Li, X.-Z. Liang, B. J. McAveny, V. P. Meleshko, J. F. B. Mitchell, J.-J. Morcrette, D. A. Randall, L Rikus, E. Roeckner, J. F. Royer, U. Schlese, D. S. Sheinin, A. Slingo, A. P. Sokolov, K. E. Taylor, W. M. Washington, R. T. Wetherald, I. Yagai, and M.-H. Zhang, Intercomparison and interpretation of climate feedback processes in 19 atmospheric general circulation models. *J. Geophys. Res.*, **95,** 16, 601-16, 615, 1990.

Cheng, L., T. C. Yip, and H. R. Cho, Determination of mean cumulus cloud vorticity from GATE A/B scale potential vorticity budget. *J. Atmos. Sci.*, **37,** 733–753, 1980.

Chimonas, G. and R. Rossi, The relationship between tropopause potential temperature and the buoyant energy of storm air. *J. Atmos. Sci.,* **44,** 2902–2911, 1987.

Chiou, E. W., M. P. McCormick, W. P. Chu, and G. K. Yue, Global distributions of cirrus determined from SAGE II occultation measurements between November 1984 and October 1988. Preprints, *Conf. Cloud Physics,* San Francisco, Amer. Meteor. Soc., 513–517, 1990.

Chou, M. D., Atmospheric solar heating rate in water vapour bands. *J. Climate Appl. Meteor.,* **25,** 1532–1542, 1986.

Chou, M. D., D. P. Krats, and W. Ridgway, Infrared radiation parameterizations in numerical climate models. *J. Climate,* **4,** 424–437, 1991.

CLIMAP, The surface of the ice-age earth. *Science,* **191,** 1131–1144, 1976.

Emanuel, K. A., A scheme for representing cumulus convection in large-scale models. *J. Atmos. Sci.,* **48,** 2313–2335, 1991.

Fairbridge, R. W., Planetary beats and geology, in *New Approaches in Geomagnetism and the Earth's Rotation*, S. Flodmark, editor, World Scientific, London, 222–244, 1991.

Gamache, J. F. and R. A. Houze, Water budget of a meso-scale convective system in the tropics. *J. Atmos. Sci.,* **40,** 1835–1850, 1983.

Geleyn, J.-F., C. Girard, and J.-F. Louis, A simple parameterization of moist convection for large-scale atmospheric models. *Beitr. Phys. Atmosph.,* **55,** 325–334, 1982.

Houghton, J. T., G. J. Jenkins, and J. J. Ephraums, Eds, *Climate Change. The IPCC Scientific Assessment.* Cambridge University Press, Cambridge, 365 pp, 1990.

Lindzen, R. S., Some remarks on cumulus parameterization. Report on NASA-GISS Workshop: *Clouds in climate-modelling and satellite observational studies,* 42–51. (N. B. The contents of this paper are largely to be found in Lindzen (1988)), 1981.

Lindzen, R. S., Some remarks on cumulus parameterization. *Pure Appl. Geophys.,* **126,** 123–135, 1988.

Lindzen, R. S., Some coolness concerning global warming. *Bull. Amer. Meteorol Soc.,* **71,** 288–299, 1990.

Lindzen, R. S., A. Y. Hou, and B. F. Farrell, The role of convective model choice in calculating the climate impact of doubling CO_2. *J. Atmos. Sci.,* **39,** 1189–1205, 1982.

McClatchey, R. A., R. W. Fenn, J. E. A. Selby, F. E. Volz, and J. S. Garing, *Optical properties of the atmosphere,* Third edition AFCRL-72-0497, 108 pp. [NTIS N7318412.], 1972.

Ogura, Y. and H. R. Cho, Diagnostic determination of cumulus cloud population from observed large-scale variables. *J. Atmos. Sci.,* **30,** 1276–1286, 1973.

Oort, A. H., Global atmospheric circulation statistics, 1958–1973, *NOAA Professional Paper,* **14,** 180 pp, 1983.

Rasch, P. J. and D. L. Williamson, Computational aspects of moisture transport in global models of the atmosphere. *Q. J. Roy. Meteorol. Soc.,* **116,** 1071–1090, 1990.

Raval, A. and V. Ramanathan, Observational determination of the greenhouse effect. *Nature,* **342,** 758–761, 1989.

Rind, D., Puzzles from the tropics. *Nature,* **346,** 317–318, 1990.

Rind, D. and D. Peteet, Terrestrial conditions at the last glacial maximum and CLIMAP sea-surface temperature estimates: Are they consistent? *Quaternary Res.* **24,** 1–22, 1985.

Rind, D., E. W. Chiou, W. Chu, J. Larsen, S. Oltmans, J. Lerner, M. P. McCormick, and L. McMaster, Positive water vapor feedback in climate models confirmed by satellite data. *Nature,* **349,** 500–503, 1991.

Rutledge, S. A., E. R. Williams, and T. D. Keenan, The down under doppler and electricity experiment: Overview and preliminary results. *Bull. Amer. Meteorol. Soc.,* **73,** 3–16, 1992.

Sarachick, E. S., Tropical sea surface temperature: an interactive one-dimensional atmosphere-ocean model. *Dyn. Atmos. Oceans,* **2,** 455–469, 1978.

Sarachick, E. S., A simple theory for the vertical structure of the tropical atmosphere. *Pure Appl. Geophys.,* **123,** 261–271, 1985.

Schneider, E. K., Axially symmetric steady-state models of the basic state for instability and climate studies. Part II: Nonlinear calculations, *J. Atmos. Sci.,* **34,** 280–296, 1977.

Sun, D.-Z., *Tropical tropospheric water vapor budget, maintenance of the lapse rate, and distribution of the extratropical tropospheric temperature and wind.* Ph.D. Thesis, M.I.T., 203 pp, 1992.

Sun, D.-Z. and R. S. Lindzen, On the distribution of tropical tropospheric water vapor. *J. Atmos. Sci.,* to appear, 1992.

Szoke, E. J., E. J. Zipser, and D. P. Jorgenson, A radar study of convective cells in meso-scale systems in GATE, I: vertical profile statistics and comparisons with hurricanes. *J. Atmos. Sci.,* **43,** 182–197, 1986.

Williams, E. R., S. A. Rutledge, S. G. Geotis, N. Renno, E. Rasmussen, and T. Rickenbach, A radar and electrical study of tropical "hot towers". *J. Atmos. Sci.,* **19,** 1386–1395, 1992.