Subtropical Water Vapor
As a Mediator of Rapid Global Climate Change

Raymond T. Pierrehumbert
University of Chicago,Chicago, IL

Abstract

This article surveys the essential features of atmospheric water vapor dynamics needed to address current issues regarding the possible role of water vapor changes in mediating climate fluctuations on the millennial to Milankovic time scales. The focus is on the subtropics, which afford the most interesting possibilities for significant feedbacks. The observed distribution of water vapor, the amount by which water vapor must change in order to cause a significant temperature change, and the physical factors that determine the water vapor content of the subtropical atmosphere are discussed. It is shown that halving the subtropical relative humidity would lead to a 2.5K cooling of the tropics, while doubling it would lead to a 3K warming. The humidity content of the subtropics could be reduced by enhancing subsidence, reducing transient eddy activity, or contracting the convective region. Further work is needed to determine which, if any, of these changes occur in concert with the observed millennial and longer scale climate fluctions.

1. INTRODUCTION

High resolution climate proxy records covering the past few hundred thousand years have revealed that, embedded within the long-period Milankovic scale fluctuations are a bewildering variety of persistent shorter period fluctuations without any obvious tie to an external forcing mechanism of the climate system. Climate variations of this sort have shown up in polar and tropical ice cores, corals, and ocean sediment cores. Geochemical signatures of temperature and precipitation variations have been found in alkenones, planktonic and benthic foraminifera $\delta^{18}O$ and $\delta^{13}C$, glacial $\delta^{18}O$ and δD, foramanifera assemblages, dust, ice-rafted debris and methane, to name a few. Climate variations often involve rapid transitions to vastly different states, and can involve time scales from decades to centuries to millenia and on upward. Indeed, it is not at all clear that there is any real spectral gap between what we think of as "weather" and the ponderous 100,000 year waking and sleeping of the great ice sheets. Certain millennial scale events have names, including the Dansgaard-Oeschger events, the Heinrich events and the Bond cycles; myriad others still await christening. Milankovic-scale cycles exhibit a fair amount of global synchrony, and the globalization effect of CO_2 fluctuations can probably account for a good bit of this synchrony. The shorter period events do not exhibit such a clear and consistent spatial pattern, but there is considerable evidence that they do not represent purely local phenomena. They are evidently part of a loosely coupled global system. Evidence of this nature is presented in numerous contributions elsewhere in the present volume.

The best studied of candidate mechanisms for millennial scale climate fluctuations involves the switching on and off of North Atlantic Deep Water formation. However, as reviewed by [Broecker (1994)], and elsewhere in the present volume, it is not entirely clear that the mechanism can yield a sufficiently strong response outside the North Atlantic region to account for the observations. General circulation model evidence suggests that with presently understood physics, even a change as major as introducing a Northern Hemisphere ice sheet may induce only a weak response in the Northern Hemisphere tropics, and hardly any response at all south of the Equator [Manabe and Broccoli (1985)]. It is not certain that the millennial scale fluctuations observed in different parts of the globe are all related to each other, but it is certain that such fluctuations are observed just about everywhere, so one should at least entertain the possibility that the NADW picture alone cannot account for everything that is going on. One either needs independent fluctuation mechanisms in the various regions, or a means of globalizing the impact of the NADW signal, or some combination of the two.

Carbon dioxide, methane and N_2O do not seem to exhibit fluctuations of sufficient amplitude to couple the different regions on millennial time scales. In view of the strong jet streams, atmospheric transports alone are probably sufficient to spread the influence of a strong regional event over the entire extratropical hemisphere in which it occurs. There is evidence that a combination of atmospheric and oceanic heat transports can extend the effect of a Northern extratropical anomaly at least part way into the tropics. Almost any change in the circulation induced will have repercussions for atmospheric water vapor. Water vapor has a rapid response time, and because it it such a good greenhouse gas, it has a large leverage over climate. The tropics sit in a good place to couple climate changes between the Northern and Southern hemispheres. Further, because tropical climate is dynamically constrained to be quite zonally uniform, the tropics provide a natural mechanism for spreading climate effects across longitudes. Therefore, it is natural that thoughts should turn in the direction of tropical water vapor when faced with the necessity of finding mechanisms for globalizing rapid climate change. One is on the lookout especially for "threshold" effects which could switch on or off suddenly, and give rise to rapid changes of climate.

This paper surveys the basic aspects of water vapor and climate that provide essential background for addressing questions regarding the role of the atmosphere in millennial scale climate change. Most of the subject matter is also relevant to the general problem of cooling of the Tropics during glacial times. With regard to mechanisms, we will be content to point out some points of vulnerability of climate, which are candidates for amplifying and globalizing climate variations. Our focus is on the subtropics, which exhibit a rich interplay of dynamics, thermodynamics and radiative effects offering much scope for interesting transitions. We will be concerned with moisture in the free atmosphere rather than boundary layer; free-atmosphere moisture has a sharp gradient between the subtropics and the convective region, whereas boundary layer moisture is very uniform. Further, it is the free-atmosphere water vapor which has the most leverage over the radiation budget of the planet.

Section 2 provides an overview of the observed tropical humidity pattern, and of the basic workings of tropical climate. The impact of water vapor changes on tropical temperature and circulation is discussed in Section 3. Mechanisms determining the subtropical water vapor content, and ways in which they may change in a changing climate, are laid out in Section 4. In Section 5 a few remarks are offered as to where water vapor may fit into the bigger picture of millennial climate fluctuation, and on the rocks that need to be turned over in the search for a theory of these enigmatic events.

2. THE OBSERVED HUMIDITY PATTERN

Dynamically, the Tropics is the region of the Earth that lies under the influence of the large scale overturning Hadley and Walker circulations. Roughly speaking, it extends from latitudes 30S to 30N. The division of the tropics into convective (ascending) and nonconvective (subsiding) regions is crucial to the understanding of tropical climate in general, and tropical water vapor feedbacks in particular. This situation contrasts with the extratropics; there, mobile transient baroclinic eddies dominate the climate, and large scale geographically fixed regions of preferred ascent or descent do not play a prominent role. Throughout the following, we will interchangeably refer to the "non-convective region" of the tropics as the "subsiding region," "the subtropics," or "the dry pool."

Only a small portion of the tropics is subject to direct moistening by nearby deep convection. This can be seen clearly from the climatological precipitation pattern, a typical example of which is shown in Plate 1. In the tropics, heavy precipitation (a marker of deep convection) is confined to three zones: the narrow zonally oriented ITCZ band, the Pacific warm pool region, and continental rainfall regions in S. America and Africa. These zones make up about one third of the area of the tropics. The rest of the tropical free atmosphere must get its moisture by some more indirect means.

The convective region is characterized by mean ascent, with the resulting adiabatic cooling balancing the latent heat release due to precipitation. The nonconvective region is characterized by mean descent, with the resulting adiabatic warming balancing infrared radiative cooling. We will amplify on this picture in Section 3. For now, the key point to keep in mind is that, in the absence of a compensating moisture source, subsidence would cause the subtropical water vapor mixing ratio to relax to the tropopause value - around 10^{-5} - over the course of about 30 days. This relaxation occurs because mixing ratio (also known as specific humidity) is conserved following parcel trajectories. The subsidence thus brings down arid air from the tropopause level, and this air is replaced by outflow from the convective region, which, for consistency with this picture, must also be at the tropopause level. The observed subtropics are dry, but not so dry as would be indicated by this process, so there is evidently a pathway for moist air to reach the subtropics without being wrung out through the cold tropopause.

In Plate 2 we show the pattern of 500-700mb mean moisture mixing ratio for the period March 24-28, 1993. This field was reconstructed from TOVS polar orbiter satellite data. There is a close association between tropical moist regions and regions of high climatological precipitation, even though the moisture data is from a single five-day period. The moist regions generally spread somewhat beyond the high precipitation zones, particularly in the Southern hemisphere tropics. To the north of the ITCZ, there is a very sharp boundary between moist, convective air and the dry pool, both in the Atlantic and in the Pacific. The dry subtropical air in places has specific humidity below .001, but more typically has specific humidity between .001 and .002 at the 500-700mb level. Air with a mixing ratio of .001 would be saturated at the 280mb level, and presumably originated in outflow from the convective region near that level before subsiding to the layer analyzed in Plate 2. Air with a mixing ratio of .0015 would originate from near the 305mb level. The subtropical air is highly undersaturated, with relative humidities in the range of 10%-15%, with occasional values even drier. We have examined the corresponding TOVS data up to 200mb, and found that the mixing ratio is not constant with height in the subtropics, as would be expected if all the outflow were at a single high-altitude level. Instead, the mixing ratio decreases gently with height, in a manner suggestive of moisture injection at multiple levels. The vertical profile of mixing ratio is intermediate between a constant mixing-ratio model and a constant relative-humidity model.

It is important to recognize that the amount of water required to moisten the subtropics is truly tiny. With a mixing ratio of .001 throughout the 500-700mb level, the water content of this layer is equivalent to a layer of only 2 mm of liquid water. Higher layers contain even less water. If we take a typical rainfall rate of 300 mm/month and divide by three to account for the fact that only about a third of the area of the tropics experiences significant precipitation, we find that only 2% of the moisture in the rainfall needs to be retained in the atmosphere over the course of a month in order to achieve the observed moisture level. Clearly, water supply is not the limiting factor in subtropical water vapor. Thus, it would be incorrect to infer any relation between changes in tropical precipitation rate, and changes in the subtropical moisture content.

To keep the discussion firmly grounded in reality, we will make illustrative use of four E. Pacific subtropical soundings obtained during the CEPEX experiment (Williams 1993). CEPEX was a convective-region experiment, but fortunately these four "leftover" sondes were released into the interesting subtropical region on the homeward voyage of the Vickers ship. Their positions are shown in Plates 1 and 2, and they were released at 00Z on March 24, 25, 26 and 27, proceeding from southernmost to northermost. One can not base a comprehensive picture on just four soundings, and what we shall present does not aspire to do so. Nonetheless, the nature of the tropical climate is so clearly drawn that all the "dry pool" concepts that we need to deal with are abundantly revealed in these soundings.

The temperature soundings are shown in Figure 1, and the relative humidity profiles are shown in Figure 2. In accordance with common practice, the relative humidities are reported relative to saturation over liquid , even at altitudes where the dominant condensate would be ice. Because of the limitations of the humidity sensor, the values reported above 300mb are of dubious validity, and should probably be ignored. A comprehensive discussion of the sensor, and of humidity soundings during CEPEX (apart from these four subtropical soundings) can be found in [Kley *et al* (1997)].

Although the soundings span approximately 20 degrees of latitude, the thermal structure shows very little variation amongst them. The farthest north is only slightly cooler than the one closest to the Equator, and all have the tropopause at very nearly the same altitude. This uniformity illustrates the familiar point that, above the boundary layer, the Tropics cannot support large horizontal temperature gradients. The reasons for temperature uniformity in the Tropics are part of well-established tropical lore, which one can find discussed in [Held and Hou (1980)], [Emanuel (1995)], and [Pierrehumbert (1995)]. Because the thermal structure is so tightly coupled across the whole Tropics, one can make considerable progress in understanding tropical climate by employing radiation budgets averaged over the entire tropics while ignoring regional temperature variability.

Along with each temperature sounding, we also show the saturated moist adiabat corresponding to the observed 700mb temperature. It is significant that the temperature profiles closely follow the moist adiabat even though the air is highly undersaturated, and even though the region does not experience convection. It was established by [Xu and Emanuel (1989)] that this principal holds throughout a comprehensive database of tropical soundings. Ultimately, the heating that maintains the moist adiabat in the subtropics is due to condensation, but not to local condensation; the effects of remote condensation in the convective region are transmitted to the subtropics via their effect on subsidence.

The moisture soundings underscore the general dryness of the subtropics, with relative humidities in the 800-300mb level frequently falling to values as low as 5%. There is a very sharp distinction between the moist boundary layer air (with relative humidity as high as 80%) and the arid air aloft. This generally dry background is interrupted by a number of spikes of relatively moist air. Though some spikes are very thin in the vertical,they can sometimes be thicker, as in the March 27 sounding which has a broad moist intrusion centered on 500mb. The complex vertical structure of the humidity suggests the primacy of horizontal advection as a moisture source for the subtropics. Differential advection in the vertical can create such structures by replacing a thin layer of dry subtropical air with moist air of convective origin, without disturbing the other layers.

This anecdotal picture of the extent of the dry pools, and their degree of dryness, is entirely compatible with the more extensive analysis presented by [Spencer and Braswell (1997)]. A recurrent worry in such endeavors is the reliability of radiosonde moisture data at the low values needed to properly characterize the subtropics. Reliable, comprehensive satellite-based moisture data is only now becoming widely available. The characterization of subtropical moisture will improve in the future, but it will take some years to build up an accurate picture of the climatology.

Although the free-atmosphere subtropical air is arid, the boundary layer air is moist both in the convective and nonconvective regions, with relative humidities in the vicinity of 60-70% (see, e.g. [Broecker (1997)]). Keeping this in mind, we will employ two idealized models of of the subtropical tropospheric humidity profile in subsequent discussion. In both models, the boundary layer air (below 850mb) is kept at fixed and high relative humidity. In the "RH model," we specify the the moisture profile so as to maintain fixed relative humidity in the rest of the troposphere. In the "outflow model" we instead specify a fixed specific humidity q above the boundary layer, with the exception that the humidity is set to saturation for tropospheric altitudes greater than the altitude z_s where the given specific humidity becomes saturated. This model corresponds to saturated convective region outflow at altitudes above z_s only, with moisture communicated to the rest of the subtropical troposphere only through subsidence.

3. CLIMATIC IMPACT OF SUBTROPICAL HUMIDITY

Because water vapor is a potent greenhouse gas, and because the thermal coupling in the tropics is so tight that the temperature changes in concert throughout the tropics, the dryness of the subtropics has a profound cooling effect on the whole tropics. The importance of the subtropics in climate was starkly revealed in the simplified models pursued in [Pierrehumbert (1995)]. Insofar as the tropics are about half the area of the planet, and insofar as they provide a boundary condition for the extratropics, the influence can be expected to extend to the rest of the planet. Curiously, the issue of the water vapor content of the subtropics, and the corresponding climate sensitivity noted in [Pierrehumbert (1995)], has been used in support of the claim that general circulation models are at risk of overpredicting the response to the doubling of CO_2 [Spencer (1997); Spencer and Braswell (1997)]. It is not as widely appreciated that there is an equal or greater scope for errors in the direction of *underestimating* the warming. For example, the water vapor feedback enhancing climate change would be stronger than current thinking suggests, if the dry pools were to contract in area in response to a warming planet. Given that the factors governing the dry pool area are not at all understood, this possibility must give us pause.

Be that as it may, it is the very sensitivity of climate to subtropical water vapor that brings the matter to our attention in the context of millennial scale climate change. The questions we seek to answer in this section are: How do the subtropical humidity content and radiation budget affect the tropical circulation? How much do we have to reduce the subtropical humidity to significantly cool the tropics (all other things being equal)? How much do we need to increase the subtropical humidity to substantially warm the tropics?

3.1. Radiative cooling and tropical circulations

Climate is more than just temperature, and it should be kept in mind that extra-tropical climate forcings could have a more profound impact on tropical circulations than on the tropical temperature itself. This point is especially important to the interpretation of ice-core methane records, which probably reflect changes in precipitation over tropical land areas at least as much as they reflect tropical temperature. Our main preoccupation in this Section will be with the strength of the subtropical subsidence, because the subsidence strength provides a convenient measure of the intensity of tropical circulations, and because it will turn out to be one of the key players in determining the subtropical humidity content.

In order to fully understand the effects of greenhouse gases on climate, it is essential to keep in mind that, under clear sky conditions, the troposphere of a planet generally experiences infrared radiative cooling throughout its depth; in equilibrium this cooling must be balanced by a compensating warming term, and the nature of the balance in large measure determines the character of the climate response. Under some common circumstances, increasing the concentration of a greenhouse gas intensifies the infrared cooling rate of the troposphere. This point is illustrated in Figure 3, where we show the upper tropospheric radiative cooling rate as a function of tropospheric humidity content for the March 24 temperature sounding. These calculations were performed using the NCAR radiation model [Kiehl and Briegleb (1992)] under clear sky conditions, with stratospheric and boundary layer (below 850mb) humidity held fixed at the observed values. For the rest of the troposphere, the observed moisture profile was replaced by a series of idealized profiles corresponding either to the RH model or the outflow model. To make it easier to compare results between the two moisture models, we have plotted the radiative cooling against the 700mb specific humidity for both models. A specific humidity of .001 corresponds to a relative humidity of 11% at 700mb. For both models the radiative cooling becomes stronger as humidity is increased. The two models yield somewhat different patterns of sensitivity to moisture, but in neither case is the sensitivity extreme. In the RH model, it is necessary to vary moisture by a factor of 100, between .0001 and .01, to change the radiative cooling rate from .8K/day to 1.2K/day. For comparison, the cooling rates computed using the actual observed humidity for all four sondes are: .94K/day for the 24th, .91K/day on the 25th, .95K/day on the 26th and .92K/day on the 27th. The uniformity of these radiative cooling rates suggests there may be some profit in making simplified models of the subtropics based on uniform subsidence. The values can be matched with a 700mb specific humidity of .001 in the RH model, or .003 in the outflow model.

The factors governing the cooling rate are subtle, and dependent on the vertical structure of temperature and the greenhouse gas. This importance of the latter is evident from the comparison of the RH with the outflow models. A further appreciation of the subtlety can be obtained from looking at the effects of CO_2 changes on radiative cooling, since this gas unlike water vapor has a constant mixing ratio throughout the whole atmosphere. In Figure 3 we show radiative cooling rates for the RH moisture model with 10x present CO_2 (3300 ppmv) and 1/10 present CO_2 (33ppmv). This is an extreme variation, corresponding at the high end to levels believed to have prevailed during the Cretaceous, and at the low end to values far lower than have ever occurred on Earth (noting that glacial-age CO_2 only falls to 180ppmv). Despite the large change, the upper tropospheric radiative cooling hardly changes at all, and the sign of the effect of CO_2 actually varies with the humidity. We suggest that the enhanced tropospheric radiative cooling rate typically found in doubled CO_2 climate simulations primarily results from the increase in water vapor that occurs as the climate warms, and through indirect effects resulting from the rise in temperature. It is worth noting, though, that in the upper stratosphere, elevated CO_2 unambiguously leads to intensified radiative cooling [Ramaswamy *et al.* (1996)], and in fact leads to an actual temperature drop of the upper stratosphere.

A shortcoming of the calculation shown in Figure 3 is that equilibrium is not maintained as humidity or CO_2 is increased. As the greenhouse gas concentration increases or decreases, the atmosphere would have to warm or cool to accomodate the change in infrared trapping, whereas the calculations were carried out with fixed temperature. Increasing the temperature on the one hand increases the infrared emission, which enhances radiative cooling. On the other hand, it increases the infrared flux upwelling from warm layers nearer the surface, which upon absorption would decrease the radiative cooling. *A priori*, it is not clear which effect wins, and the answer can be subtly dependent on the temperature and greenhouse gas profile. Fortunately, an elementary result can be obtained which links the mean tropospheric cooling rate to the radiation balance at the surface and at the top of the atmosphere. Let $F(p) = I^+ - I^-$ be net net infrared radiation flux (upward positive). Then, the infrared heating rate in degrees K per unit time is

$$H_{IR} = \frac{g}{c_p}\frac{d}{dp}F(p) \qquad (1)$$

where c_p is the specific heat of air at constant pressure and g is the acceleration of gravity. Then, integrating over the depth of the troposphere yields

$$\begin{aligned}\bar{H}_{IR} &\equiv \frac{1}{p_s - p_t}\int_{p_t}^{p_s} H_{IR}dp \qquad (2)\\ &= \frac{g}{(p_s - p_t)c_p}F(p_s) - \frac{g}{(p_s - p_t)c_p}F(p_t)\end{aligned}$$

where p_s is the surface pressure and p_t is the tropopause pressure. If the stratosphere is optically thin in the infrared, then $F(p_t)$ is approximately equal to $I^+(p = 0)$, which is also known as "outgoing longwave radiation," abbreviated "OLR". The OLR is positive everywhere, since there is no significant infrared source from space, and its average over the whole planet must equal the absorbed solar radiation, insofar as the planet as a whole is nearly in equilibrium. The remaining term $F(p_s)$ in Eqn. (2) is the net infrared imbalance at the surface. It tends to be small for the following reason. Turbulent fluxes of evaporative and sensible heat keep the surface temperature rather close to the overlying air temperature. Further, the high water vapor content of boundary layer air means that the boundary layer is optically thick in the infrared, and hence radiates back to the surface most of the infrared welling upward from the surface. If this balance were perfect, $F(p_s)$ would vanish, and the mean infrared cooling of the troposphere would be simply the OLR divided by the heat capacity of the troposphere per unit area. Except in the rare circumstances where $F(p_s)$ is not small, the OLR term dominates Eqn. (2) and the troposphere experiences a net infrared radiative cooling, simply because infrared energy drains out the top of the atmosphere faster than it is replenished at the bottom. It is the turbulent coupling of the surface to the overlying air that supplies the necessary ingredient for the latter part of this general argument. As an example of the application of Eqn. (2), an OLR of 240 W/m^2 and a surface cooling of 50 W/m^2 yields a mean tropospheric radiative cooling of 2K/day if p_t is taken to be 200mb. This whole-troposphere average cooling is distinctly greater than the *mid-tropopospheric* cooling directly computed for the CEPEX soundings, because the cooling rate tends to increase sharply toward the ground.

The behavior of the infrared radiative cooling depends intimately on what is going on at the surface of the planet, and so the surface budget bears closer scrutiny. The equilibrium surface budget is

$$F(p_s) + E + kc_p(T_s - T_a) = S(p_s) + F_o \quad (3)$$

where E is the evaporative heat flux out of the surface, k is the wind-dependent sensible heat transfer coefficient, T_s is the surface temperature, T_a is the low-level air temperature, F_o is the implied heat flux due to ocean currents (set to zero over land), and $S(p_s)$ is the solar radiation flux absorbed at the surface. If the right hand side of Eqn.(3) is held fixed, the combination of evaporative and sensible heat flux must go up as $F(p_s)$ goes down, and conversely. Except over dry, hot deserts, the evaporation dominates the sensible heat flux in the warmer parts of the world. As $F(p_s)$ approaches zero, the evaporation dominantly balances the surface solar heating corrected for oceanic internal fluxes; the latter allow the evaporation to greatly exceed locally absorbed solar radiation over the Gulf Stream, for example.

Using the same Legates/MSU precipitation climatology shown for March in Plate 1, it is found that the global mean precipitation at present is .098 meters per month, corresponding to an evaporative flux of 94.5 W/m^2, which is already nearly half of the roughly 200 W/m^2 of solar radiation which is absorbed globally at the surface. In the tropics, the precipitation is equivalent to an evaporative flux of 111 W/m^2, which accounts for an even higher proportion of solar radiation reaching the surface, leaving typical values of $F(p_s)$ of 60 W/m^2 or less. Hence there is rather limited scope for increasing the tropospheric radiative cooling in a warmer world. Given typical subtropical OLR values of 240 W/m^2, the tropospheric radiative cooling would only increase 25% even if $F(p_s)$ were zeroed out completely.

The general increase of precipitation with temperature discussed in the preceding paragraph is at the heart of the CO_2 thermostat regulating the temperature of the planet over geological time scales [Berner (1994)], since removal of atmospheric CO_2 by weathering into carbonate rocks is proportional to precipitation rate over land. In thinking about this thermostat, it would be well to keep in mind the limitations on precipitation increase imposed by Eqn. (3). There is no such intrinsic limitation on the *cooling* side however. In a sufficiently cold (or dry) world, $F(p_s)$ could be increased until the tropospheric radiative cooling were reduced to the point where it balances the sensible heat flux term in Eqn. (3), and precipitation is eliminated entirely.

The preceding should impart some intuition as to the circumstances under which tropospheric radiative cooling intensifies. What are the implications for the circulation of the tropics? The fact that increasing the greenhouse gas concentration increases atmospheric radiative cooling may seem paradoxical. How can the atmosphere warm in the face of a cooling term which increases in magnitude? The atmosphere can warm despite increased radiative cooling, because this cooling is balanced by other heating terms. In the tropics, the compensating heating is provided primarily by adiabatic compression due to subsidence (in the nonconvective regions) or by latent heat release (in the convective regions), as depicted in Figure 4. Though there are additional heating and cooling terms, including heating due to atmospheric absorption of solar radiation, and cooling due to import of cold extratropical air by eddies, the idealization shown in Figure 4 will serve us for a while. The dominant balance in the nonconvective region may be expressed

$$c_p\omega\frac{\partial\theta}{\partial p} = H_{IR} \tag{4}$$

where ω is the rate of change of pressure following a fluid parcel and θ is the dry potential temperature. Hence, the subsidence rate as measured by ω depends on the stratification of the atmosphere. Even if H_{IR} is held fixed, the subsidence rate becomes unbounded as the atmosphere approaches the dry adiabat $\theta = const.$ As the atmosphere becomes colder and drier, water vapor has progressively less effect on stratification and one might indeed expect the dry adiabat to be approached. However, over the range of temperatures of interest in the present essay, the effect of stratification on subsidence is not pronounced. Let us assume the atmosphere to lie on the moist adiabat, and measure the stratification by $\theta_{500-700} \equiv \theta(500mb) - \theta(700mb)$. For a modern surface temperature of 300K, we find $\theta_{500-700} = 17.1K$. Reducing the surface temperature to a plausible glacial temperature of 295K only drops $\theta_{500-700}$ to 14.4K, and going to an extreme cooling of 290K drops $\theta_{500-700}$ to 12K. In order to double the subsidence with fixed radiative cooling, it is necessary to drop the surface temperature all the way to 282.5K.

The solar-powered compressor depicted in Figure 4 can operate even if there is no top-of-atmosphere radiative imbalance (i.e. if the absorbed solar radiation S equals the OLR locally), nor does it require a meridional gradient in either S or OLR. Consider the idealized case where the surface sensible heating vanishes and $F(p_s) = 0$, and further suppose the atmosphere to be perfectly transparent to solar radiation. Then, there is subsidence at a rate proportional to the OLR, while S independently penetrates to the surface and is balanced by evaporation; the evaporated moisture is carried away by the current fed by the subsidence, maintaining a steady state. In some sense, the compressor lives off the blackbody temperature difference between the "hot" incoming solar radiation and the "cold" OLR. Yet, for *identical* top-of-atmosphere and surface energy budgets, the system can also support a solution in local radiative-convective equilibrium, with no large scale circulation. It is for this reason that it is difficult to make simple and sweeping generalizations as to the circumstances in which the circulation gets stronger, or weaker.

In the *convective* region the infrared cooling would balance the latent heat release due to precipitation, if there were no mean ascent. With a large scale circulation such as is necessitated by compatibility with the downward mass flux in the subsiding region moisture is imported into the convective region, and the latent heat release more than balances the convective-region radiative cooling, allowing a mean ascent to prevail there. Some of the dry air descending in the subsiding region infiltrates the boundary layer, which keeps the boundary layer unsaturated and sustains the evaporation and lateral moisture flux needed to feed the extra latent heat release needed in the ascending region. The upshot is that increased subtropical radiative cooling has the effect of increasing the strength of the Hadley and Walker circulations, which leads to enhanced precipitation in the convective region (so long as the boundary layer humidity doesn't go down too much). Conversely, reduction in the subtropical radiative cooling is expected to lead to a weakened circulation and reduced precipitation.

The tropospheric radiative cooling at a given site can be changed either by changing the OLR or by changing $F(p_s)$. If there is no change in the energy exported from the atmosphere-ocean column, then OLR can change only if the local solar absorption is changed, e.g. by increasing the albedo through increasing cloud cover (which, in the subsiding region, could be accomplished through increase in low cloud cover, accompanied by a cooling of the subtropics and consequent reduction in OLR). The surface infrared cooling, $F(p_s)$ can be changed by altering conditions at the surface, even if OLR is kept fixed.

In the tropics, $F(p_s)$ is particularly small, because evaporative coupling of the surface to the boundary layer air is strong, and the water vapor content of the boundary layer is high. If the tropics is made warmer by any means whatsoever, then the boundary layer gets moister, $F(p_s)$ tends to get smaller, and evaporative coupling tends to get stronger,. Unless the OLR decreases significantly, the infrared radiative cooling of the troposphere must then increase; according to Eqn. (3) the evaporation, and hence precipitation and latent heat release, increase by a corresponding amount. Conversely, in a colder world $F(p_s)$ would tend to be more competitive with evaporation, and radiative cooling, evaporation, and precipitation would all decrease. This scenario is plausible, but not inevitable. Evaporation could increase in a colder world if surface winds intensified dramatically, or boundary layer relative humidity dropped. The latter has the dual effect of increasing moisture flux for a given surface wind intensity, and also increasing the radiative cooling of the surface. Changes in low level clouds also profoundly affect the tropospheric radiative cooling. Low clouds reduce the local solar absorption, and hence, in equilibrium, reduce the OLR. They also are very effective at making the boundary layer optically thick in the infrared, which reduces $F(p_s)$. If the former effect wins, tropospheric radiative cooling decreases, whereas if the latter effect wins the tropospheric radiative cooling increases (though the influence may show up largely near the boundary layer). Despite the complicating possibilities, the naive picture does seem to have some merit, as global mean precipitation and tropospheric radiative cooling do generally increase in numerical experiments in which the world is made warmer by doubling CO_2. On the other hand, [Ramstein *et al.* (1998)] report simulations in which the Hadley cell strength *increases* for cold LGM conditions, but *decreases* for warm doubled CO_2 conditions. Also [Knutson and Manabe (1995)] report quadrupled CO_2 simulations in which the tropical Pacific precipitation increases despite a slight *weakening* of the tropical Pacific overturning circulation.

The tropospheric radiative cooling can also be affected by oceanic influences: If upwelling or horizontal advection of cold water results in the surface temperature becoming substantially less than the overlying air temperature (as happens in the cold tongue of the tropical Pacific),then $F(p_s)$ is reduced, or even becomes negative, and tropospheric radiative cooling is enhanced.

The tropics are not isolated, but rather export excess solar energy to the extratropics. How does the atmospheric part of this transport affect the subsidence? The effect is important for the purposes of this essay, because climate forcings indigenous to the Northern Hemisphere, such as ice sheets or North Atlantic temperature changes, communicate their influence to the tropics in part through changes in horizontal atmospheric transports. Atmospheric *sensible* heat transport mixes cold extratropical air into the subtropics, and acts in concert with the radiative cooling shown in Figure 4 to maintain the subsidence. However, if the sensible heat transport were increased, the OLR the tropics needs to radiate decreases by a like amount, which by Eqn. (2) reduces the subsidence so as to exactly offset the effects of increased sensible heat transport. An increase in sensible heat transport could still enhance subsidence if it were concentrated near the edge of the subtropics, as in the ice-sheet experiments of [Manabe and Broccoli (1985)]. In that case, the subtropical cooling due to admixture of cold air can locally dominate the required reduction of OLR, which is spread over the whole low latitude band. An increase in horizontal latent heat transport, on the other hand, does not lead directly to a local cooling which affects subsidence; it still reduces the tropical OLR, in equilibrium, and hence reduces subsidence. Finally, if the net lateral atmospheric heat transport is held fixed while its composition shifts from latent to sensible fluxes, then the subsidence should intensify since OLR stays fixed but subtropical internal cooling increases. In the experiments on the influence of a Northern Hemisphere ice sheet by [Manabe and Broccoli (1985)], the net atmospheric heat transport out of the tropics increases only slightly, while there is a large shift from latent to sensible heat transport, in accord with the general reduction of water vapor content in the colder atmosphere.

To the extent that the subsiding regions are zonally symmetric, increasing the intensity of the subsidence can be taken as equivalent to "increasing the strength of the Hadley circulation." The very presence of a Hadley circulation of any type redistributes heat within the low latitudes, making the equatorial regions cooler, and the subtropical regions warmer, than they would be if the low latitudes were in local radiative-convective equilibrium. However, increasing the strength of the Hadley circulation does not have a marked effect on the tropical temperature distribution. So long as the circulation is strong enough to approximately conserve angular momentum aloft in the face of frictional dissipation, the temperature converges to a universal meridional profile independent of the strength of the circulation [Held and Hou (1980)]. For related reasons, it is not easy to change the meridional extent of the Hadley circulation appreciably. There is some leeway for the circulation strength to alter the temperature pattern, since the observed subtropical jet is only about half as strong as it would be if angular momentum were conserved; hence dissipation of angular momentum is evidently not completely overwhelmed, at least toward the poleward boundaries of the subtropics. Increasing the strength of the circulation does, in general, have the following effects: (1) It strengthens the surface easterlies, since the easterlies are determined by a balance between the zonal Coriolis acceleration and surface friction, and the Coriolis acceleration is proportional to the low level meridional flow. (2) It increases the strength of the subtropical jets aloft, especially poleward of 15N or 15S, since friction has less time to dissipate angular momentum if the meridional transport is faster. (3) Thermal wind balance requires that, to the extent the subtropical jets strengthen, the meridional temperature gradient between the equator and the poleward edges of the subtropics increases. This effect would retard the cooling of the subtropics compared to that of the extratropics, and would act conversely in the event of a weakening Hadley circulation. (4) It increases the precipitation in the Intertropical Convergence Zone.

Our preoccupation with subsidence should not be construed to mean that the the Hadley circulation is "driven by" subtropical cooling, though this point of view is probably closer to the truth than the more conventional one that the circulation is "driven by" latent heat release. In fact, the whole system is coupled, and while the subtropical cooling must somehow adjust itself so as to be compatible with the changes in the circulation, it is not necessarily the prime mover in such changes. For example [Lindzen and Hou (1988)] and [Hou and Lindzen (1992)] found that concentrating heating or moving its center off the equator leads to an intensified circulation; the subtropical radiative cooling adjusts to accomodate these changes, but cannot be regarded as the *cause.* The Hadley circulation is a bit like a flywheel, which tends to stay in motion once it is set in motion. Its ultimate speed is determined by the balance between the aggregate of small dissipating and accelerating mechanisms.

3.2. Influence of water vapor on the radiation budget

To determine the atmospheric temperature change, we must look to the top-of-atmosphere radiation budget. The temperature of a planet is determined by a balance between infrared cooling to space (the "OLR"), and heating by absorption of solar radiation. Any factor that reduces OLR for fixed temperature will have a warming effect, since the planet will then have to heat up in order to bring the OLR back to the point where it can balance the solar radiation. For example, doubling the CO_2 concentration from preindustrial values reduces the Earth's OLR for fixed temperature by about 4 W/m^2, which can be compensated by a rise in temperature of 2-4C,depending on water vapor and cloud feedbacks.

The potent influence of moisture on the subtropical part of the Earth's radiation budget is seen in Figure 5, where we show the OLR as a function of moisture for typical tropical temperature profiles. The temperature profiles are from the four soundings of Figure 1, plus one additional sounding from 4S near the warm pool region. The results were computed using both the outflow-model idealized moisture profiles and the RH model. Note that for fixed humidity, the OLR decreases gently with distance away from the Equator, but that the effect of moisture variation is far more pronounced than the temperature effect. From these results we can conclude that the OLR is highly sensitive to the small amounts of water vapor contained in the dry pools; dry though they be, making them yet drier would substantially change the OLR. For example, in the outflow model, changing the specific humidity from 10^{-5} (about as dry as it could plausibly get) to 10^{-4} drops the OLR by 25 W/m^2. Increasing the moisture by another factor of 10 drops the OLR by another 25 W/m^2. In fact, except for very dry values, the OLR in the outflow model responds logarithmically to moisture, much as it is known to do for CO_2. The associated elevated sensitivity of OLR to moisture at very dry values — in the sense that adding one molecule of water to a dry atmosphere drops OLR more in absolute terms than adding the same molecule to a moist atmosphere — has been noted by [Spencer and Braswell (1997)], among others. The RH model shows an increase of sensitivity (in the logarithmic sense) for specific humidities above .001, but, recognizing that the results in Figure 5 are presented with a logarithmic humidity axis, this model too indicates high sensitivity to small changes in the humidity of dry air.

The extreme cases in [Pierrehumbert (1995)] in which increasing the dry pool greenhouse gas concentration led to reduced atmospheric temperatures were fundamentally different from the realistic situation treated above, in that the former atmospheres had temperatures far exceeding that of the surface underlying the dry pool. In this case, increasing the infrared optical thickness of the atmosphere actually increased the OLR, by replacing the cold radiating surface with a warmer one in the atmosphere. The atmosphere must then cool down in order to restore balance. This is a classic "anti-greenhouse" effect, the conditions for which cannot easily be met in any plausible climate change scenario on Earth. This does not mean that the "radiator fin" effect described in [Pierrehumbert (1995)] is irrelevant. The dryness of the subtropics still make the climate cooler than it would otherwise be, and in this sense the subtropics indeed act to help the deep tropics radiate away its excess heat. However, the Earth is almost invariably in a regime where increasing the subtropical moisture inhibits this radiation, and thus has a warming effect.

For our purposes, it can safely be assumed that adding moisture to the dry pool will make the atmosphere warmer, and removing it would make the atmosphere cooler But how much warmer or cooler? In order to answer this question, we turn to a simple model of the radiation balance of the whole tropics, comprising the convective and dry-pool regions. The assumptions are as follows: (1) The temperature is horizontally uniform over the entire tropics. (2) The vertical temperature profile is on the moist adiabat corresponding to the surface temperature (which, in turn, is continuous with the low level air temperature). (3) The convective region has 70% relative humidity throughout the troposphere. (4) The dry pool humidity is specified according to the RH model, with various specified free-tropospheric humidities. (5) The convective region occupies $\frac{1}{3}$ of the area of the tropics, while the dry pool occupies the remaining $\frac{2}{3}$ of the tropics. We will neglect the effect of clouds on the radiation budget, on the grounds that the high clouds of the present tropics have a nearly zero net effect on the radiation budget [Ramanathan *et al.* (1989)]. Whether this cancellation would continue to hold in an altered climate is anybody's guess. As discussed by [Miller (1997)], there is a possible important role for low marine stratus clouds in affecting climate change. This is an important consideration, but one which we will not take up here as our main interest is in the leverage water vapor has over climate.

We adopted the RH model for the dry pool on the grounds that if the dry pool moisture is ultimately drawn from the convection region, the dry pool humidity ought to go up with temperature roughly according to the way moisture in the convective region behaves. One could achieve a similar behavior in the outflow model by pegging the specific humidity to the value attained at a given altitude in the convective region. We present here results only for the RH model dry pool.

With the above assumptions, one can compute the average tropical OLR as a function of the surface temperature. This curve is shown in Figure 6, for various values of the dry pool relative humidity. These results were computed with CO_2 set to its pre-industrial value of 280 ppmv. If the tropics were energetically closed, then one would obtain the surface temperature by balancing OLR against the absorbed solar radiation. However, this would give an excessively high temperature, since the tropics in fact export energy to the extratropics, cooling the tropics and warming the extratropics. Since we are mainly interested in finding how temperature changes with changes in dry pool humidity, we will adopt the expedient of adjusting the energy input for a "base case" such that the surface temperature comes out to a reasonable value. The required energy input is the true absorbed solar radiation minus the energy (per square meter of tropics) exported to the extratropics. Changes can then be assessed with respect to this base case. This procedure implicitly assumes that the dynamical heat export from the tropics which occurs both in the atmosphere and ocean does not change with changing climate. This is a debatable assumption, but doing better requires a full general circulation model.

Let us take as our base case a situation with 10% dry-pool humidity, which is consistent with the observed values shown in Plate 2. The tropics is in balance at 300K surface temperature if the net absorbed energy is 312 W/m^2. With an estimated absorbed solar radiation of 360 W/m^2 over 30N-30S (adjusted for the joint effects of clouds on solar reflection and OLR), this implies that the tropics must export about 48 W/m^2 to the extratropics to come into balance. In view of the large role of the dynamical heat transport, it is to be expected that climate changes would alter the heat export, which in turn would have a strong affect on tropical climate. Although the atmosphere accounts for a large part of the required energy transfer (both in models and observations), there is evidence that the globe can tolerate quite large climate changes without incurring much change in the heat exported from the tropics. Notably, in the glacial maximum simulations of [Manabe and Broccoli (1985)] and [Broccoli and Manabe (1987)] it was found that cooling the extratropics by introducing an ice sheet (without reducing CO_2) had a relatively small effect on tropical temperatures, such effect as there was being confined to the north of the Equator. The smallness of the effect indicates that in the simulation the extratropics was drawing only a little more energy out of the tropics than in the present, despite the greatly increased pole-equator temperature gradient. A similar state of affairs holds in the simplified coupled atmosphere-ocean climate model of [Ganopolski *et al.* (1998)]. This result is quite counter-intuitive, given that in glacial times one has a greater temperature gradient and stronger storms,which together constitute all one needs for elevated heat fluxes. Evidently, the glacial-era atmosphere had a strengthened barrier to tropical/extratropical transport, perhaps in the form of a stronger subtropical jet. The heat export from the tropics increases, but not in proportion to the increase in temperature gradient. In [Manabe and Broccoli (1985)] it was found that there was indeed a substantial increase in *sensible* heat export across 30N, but that this was largely compensated by a decrease in *la-*

tent heat transport. In accordance with the discussion of the preceding subsection, this could lead to changes in the strength of the subsidence, even in the absence of appreciable tropical temperature changes. [Hewitt and Mitchell (1997)] repeated the ice-sheet experiments of [Manabe and Broccoli (1985)] with a different GCM, incorporating cloud feedbacks, and also found that Northern Hemisphere extratropical cooling influences lead to only weak tropical cooling and even weaker Southern Hemisphere cooling.

The GCM results of [Webb *et al.* (1997)] differ from the simulations cited above, in that they show a very large tropical and Southern Hemisphere in response to the introduction of Northern Hemisphere extratropical ice sheets, even while holding ocean heat transports fixed. It is noted by [Webb *et al.* (1997)] that the cloud parameterization employed makes the model more sensitive to changes in forcing than some other models. As in [Manabe and Broccoli (1985)], the heat exported from the tropics in [Webb *et al.* (1997)] increases slightly in the LGM climate, as compared to the present. The contrary was stated in [Webb *et al.* (1997)], but this turns out to have been due to a minus-sign error in the analysis of the results (R. Webb 1999, personal communication). On its own, this increased heat export would lead to some tropical cooling; the large response in [Webb *et al.* (1997)] can be viewed as an amplification of this cooling by the model's unusually strong cloud feedbacks. Given the very low resolution of the GCM used in [Webb *et al.* (1997)], which only has six full grid boxes covering the latitudes from 24N to 24S , the results of the simulation should be interpreted with caution, pending confirmation with higher resolution models with alternate cloud treatments. Cloud and convection patterns are quite sensitive to the tropical circulation, and the circulation response may be different in models which better resolve the tropical dynamics.

[Ganopolski *et al.* (1998)] did find an elevated energy export from the tropics in glacial times. The atmosphere accounts for most of the change across 30N, but by 20N the additional heat transport is mostly carried by the Northern subtropical ocean. It is an interesting question whether the atmosphere alone would be able to take up the slack if the oceanic heat transport were absent (as in a mixed layer ocean model). [Ganopolski *et al.* (1998)] found an increase of $.7x10^{15}$ watts in the oceanic heat flux across 20N. Averaged over the band 20N-20S, this yields an average heat export of 4 W/m^2 . On the basis of the OLR curve for the base case in Figure 6, this would lead to a tropical cooling of 1.5C, which is somewhat less than the mean oceanic cooling found in the simulation. We emphasize that this cooling includes the "standard" water vapor feedback, i.e. that which goes along with keeping relative humidity fixed as temperature changes. In the following, one ought to keep in mind that plausible increases in the oceanic heat transport can account for a substantial portion of the tropical cooling, even in the absence of fundamental changes in the nature of the water vapor feedback.

We are now in a position to consider the sensitivity of climate to changes in the subtropical humidity. If the dry pool humidity is halved to 5%, then the temperature drops to 297.4K. On the other hand, if the dry pool humidity is doubled to 20%, the tropics warms to 303.2K. Compared to the tropical temperature variations between interglacial and glacial times (estimated at 2-5K) these are highly significant figures. Still, the bottom line is that it is not enough to make small relative changes in the dry pool humidity in order to gain leverage over climate. One must have in hand mechanisms that can halve or double the humidity. To take more extreme cases, a drop in dry pool humidity to 1% cools the tropics to 293K, and a further drop to .1% cools the tropics to 290K. Clearly, the small amounts of humidity that exist in the subtropics are playing a significant role in keeping the planet warm, so it is important to know what makes the observed humidity hover around, say, 10% rather than 1%. In the other extreme, if the dry pools broke down completely, so that the convective region expanded to fill the entire tropics, the tropical temperature would shoot up to 320K. The geological record suggests strongly that mean tropical temperatures this high have never been achieved during the entire span of the fossil record, providing indirect evidence that the dry pools have never, in fact, collapsed.

These moisture effects occur jointly with the effect of CO_2 fluctuation. A recomputation of the values in Figure 6 shows that reducing CO_2 to the glacial value of 180ppm drops the OLR for a given temperature between 2 and 3 W/m^2, with the higher values obtaining for the very dry cases. Using the OLR curve for the 10% humidity base case, this decline would lead to a temperature drop of .75K to 1K, which is consistent with the GCM results of [Broccoli and Manabe (1987)]. These values are not negligable compared to the moisture effect, but the moisture effect is distinctly greater, and moreover moisture can change nearly instantaneously, whereas CO_2 cannot, and is moreover known not to have changed appreciably during most millennial scale fluctuations.

This model embodies a very crude picture of the way the tropical climate works, but it has the virtue of a more realistic radiative model of the dry pool than that adopted (for reasons of analytic simplicity) in [Pierrehumbert (1995)]. Two-box models of tropical climate have become popular recently, and even more realism has been pursued [Miller (1997); Clement and Seager (1998); Larson and Hartmann (1998)]. Much has been learned from these models, and there has even been the beginning of at attempt to take ocean dynamical transports into account. A major shortcoming of such models remains that none has been able to give an account of what determines the relative areas of convective and subsiding regions. A further need is to replace the various *ad hoc* assumptions regarding subtropical moisture content with something more dynamically based. On the latter score, at least, there is grounds to hope for rapid progress in the near future, given what has been learned about the dynamics of subtropical moisture.

4. WHAT CONTROLS SUBTROPICAL HUMIDITY?

Having established the sensitivity of climate to subtropical water vapor, we must now look to the question of what processes could change the dry pool humidity. Since there is no obvious moisture source *within* the subsiding region, the moisture of an air parcel there must be determined by tracking it back to its origins within the convection region.

Consider an air parcel moving around in the atmosphere. Its motion is characterized by its trajectory in three dimensional space, $\vec{r}(t)$. Along the trajectory, one can obtain the time series of temperature, $T(t)$, and of saturation mixing ratio $q_s(t)$. At this point, the trajectory can be defined by whatever process one likes, be it winds from observations, a general circulation model or some simplified stochastic scheme. We suppose that the specific humidity q of the air parcel is conserved except for two possible kinds of events: If the parcel wanders into a source region (e.g. the boundary layer), its moisture is reset to saturation corresponding to the value of $q_s(t)$ where the parcel encounters the region. Further, if the parcel wanders into a region so cold that its current humidity q exceeds the current saturation value $q_s(t)$, then q is reset to $q_s(t)$, the balance of the moisture being rained out. This model is a simplification of the real moisture dynamics, in that moisture changes due to mixing amongst moist and dry air parcels, and due to evaporation of precipitation falling into the parcel from above, are neglected. One may hope that both processes are not crucial in the non-convective region. With the above simplifications, the moisture at a given point is given by the minimum q_s encountered going backwards in time along the back-trajectory from the point, up to the most recent encounter with the source region. For any given model of the trajectory, this quantity can be computed by tracking q_s along the back-trajectory until the trajectory hits the source region, and taking the minimum over this finite stretch of the back-trajectory. Each back-trajectory is characterized by a unique value of $min(q_s)$, and it is the probability distribution of this quantity over the ensemble of trajectories that one must come to understand. Since q_s is primarily dependent on temperature, this quantity is primarily determined by the minimum temperature encountered along back-trajectories. Many interesting models of this statistic can be made, using either realistic winds or random walk models of the trajectories. Here, we shall confine ourselves to the most basic inferences regarding the statistic that can be made using qualitative reasoning.

A typical trajectory illustrating the preceding idea is shown in Figure 7. The chief consequence of this picture is that the longer it takes for a parcel to reach a given point in the subtropics, the drier it is. This is so because "older" air parcels have subsided more (and hence originated from colder and drier parts of the convective region); also because "older" air parcels have a higher probability of having been dried out by processing through cold extratropical regions [Yang and Pierrehumbert (1994)]. Weakened lateral mixing leads to a drier subtropics, as does enhanced subsidence. A larger temperature gradient between subtropics and extratropics also has a drying effect, since an air parcel does not have to be moved as far to be processed through a cold region.

Returning to the sondes of Figure 1, we can get an idea of the amount of subsidence needed to account for the observed dryness. In Figure 8 we plot the specific humidity q of the four sondes against the dry potential temperature θ. The potential temperature is used as the vertical coordinate because it would be conserved if there were no diabatic cooling (i.e. subsidence). In Figure 8 we also show the humidity profile $q_s(\theta)$ for the nominal "source air," which we model as the saturated mixing ratio for an equatorial sonde. If the moisture were mixed into the subtropics without subsidence, all the observed $q(\theta)$ profiles would relax to $q_s(\theta)$. Subsidence causes the air to descend to lower values of θ, which are closer to the surface. By comparing the observed $q(\theta)$ with $q_s(\theta)$, we can estimate the subsidence if the moisture originated in an air mass with humidity profile $q_s(\theta)$. For example, the moist spike at 320K in the March 24 sounding would have to have subsided from the 340K level to the 320K level, if it began life as a saturated parcel in the convective region. A similar conclusion applies to the spike at 320K in the March 27 sounding. Based on the typical cooling rates of about 1K per day indicated in Figure 3, the profiles suggest that this air took 20 days to arrive at its ultimate subtropical location. These mixing times are compatible with estimates based on direct trajectory calculations [Pierrehumbert (1998)]. The moisture variability in the soundings can be ascribed to the differences in the time it takes for the convective air to arrive at the sonde locations. "Older" air will be drier, because it has subsided more.

There is a great deal of observational support for the notion that the subtropical humidity is governed by a balance between subsidence and lateral mixing due to large-scale wind fields, where "large-scale" in this context means winds of the scales typically resolved by general circulation models [Sherwood (1996); Salathe and Hartmann (1997); Soden (1998); Pierrehumbert (1998)]. Recently, [Pierrehumbert and Roca (1998)] showed that the observed subtropical humidity pattern can be reproduced in great detail using the large scale advection-subsidence model, but that the humidity is crucially dependent on the amount of transient eddy activity. In the extreme case where transients are suppressed altogether, the subtropics become essentially completely dry. Thus, changes in the subtropical transient eddy activity, or changes in the mean flow environment which modulates the mixing action of the transients, can lead to profound changes in the subtropical moisture. Further, general circulation models must be able to faithfully reproduce the mixing action of low latitude transient eddies, if they are to accurately simulate water vapor changes.

Another way to change the water vapor content of the low latitudes is to change the area covered by the highly saturated convective region. The factors that govern the configuration of the Pacific warm pool and cold tongue are subtle and sensitive to details of the interaction between the tropical oceanic transports and atmospheric convection [Dijkstra and Neelin (1995)]. Because a small change in surface temperature relative to the air temperature can change a region from convective to nonconvective, the system governing the warm pool provides good scope for threshold phenomena, and possible drastic reorganization of convection in response to gradual climate forcings like the precessional insolation cycle.

Contemporary fluctuations in tropical convective area provide some indication of the extent to which convection can expand or contract. The convective region area, shown in Figure 9 undergoes a pronounced seasonal cycle, with peaks usually occurring in January, during the Southern Hemisphere summer. The effect of El Niño on this cycle is not completely consistent, as can be seen by comparing the convection time series with the Niño3 index time series in Figure 9. Strong El Niño's, like 1983 and 1998, tend to contract the convective region, whereas strong La Niña's,notably 1999, expand the convective area. The association is disrupted during the series of amorphous events in 1990-1995. Overall, the variations in the convective region are quite significant, ranging from 27% convective region coverage in 1998 to 34% in 1999. The indication is that a "permanent El Niño," corresponding to a breakdown of the warm pool, would yield a more zonally symmetric circulation with a smaller convective area and larger dry pool, and should lead to a cooler tropics once equilibrium is established. This hypothesized behavior conflicts with the observation that El Niño years tend to have warmer global mean surface temperatures; the discrepancy may be a matter of time scale for equilibration, or may be a matter of competing effects which overwhelm the cooling tendency of the contraction of the convective region during El Niño years.

The supposition that increasing the area of the convective region would have a warming effect (and conversely for contraction) would be straightforward if one only had to worry about water-vapor feedbacks and not cloud feedbacks. High clouds in the convective region reduce OLR beyond the water vapor effects, but also reduce solar accumulation through their albedo effects. At present, the state of affairs is such that the convective region is a strong net accumulator of energy, compared to the subsiding regions, because the cloud and water vapor OLR effects combine to dominate the albedo effects in the convective region. Any increase in the cloud albedo effect relative to the cloud OLR effect could alter this picture, and make the convective regions a cooling influence rather than a warming influence.

The four chief influences determining subtropical water vapor are thus:

Convective region humidity. All other things being equal, the subtropical moisture mixing ratio is proportional to the mixing ratio in the convective region, which is the proximate moisture source for the subtropics. It is a reasonable working hypothesis that vigorous deep convection maintains the convective region humidity at a roughly fixed fraction of its saturation value at each altitude. In consequence, a cooler climate will have a drier subtropics, in proportion to the reduction of saturation vapor pressure. This is the *conventional water vapor feedback*.

Subtropical subsidence rate. All other things being equal, a more rapid subidence rate brings down moisture from higher levels in the convective source region, and results in a drier subtropics. Since the source region mixing ratio decreases nearly exponentially with altitude, the subtropical humidity content decreases approximately exponentially with increasing subsidence.

Large scale horizontal mixing. Reduction in horizontal mixing, either by reducing transient eddy intensity, or by increasing mixing barriers by intensifying subtropical jets, reduces subtropical moisture, and leads to a cooling.

Change in area of convective region. Contracting the area of the convective regions will generally have a cooling effect, whereas expanding them would have a warming effect, provided cloud feedbacks do not change. However, changes in the balance between high cloud effects on OLR and albedo have the potential to fundamentally alter the impact of the convective regions on the radiation budget, either in the direction of warming or cooling.

In discussions of water vapor feedback, it is both appealing and common to assume that increases in evaporation imply increased atmospheric water vapor content. This assumption is a fallacy. In equilbrium, the evaporation equals the precipitation, and both represent the rate of flux of water through the system. The rapidity with which water cycles through the system has little bearing on how much vapor is retained in the system. A steady drip can fill up a bucket if the bucket is tightly caulked, but a torrent won't suffice to fill up a sieve. Likewise, a low evaporation rate can keep an atmosphere near saturation in equilibrium if the factors producing dry air are weak, and conversely a high evaporation rate can coexist with a dry atmosphere if dry air production is strong. As a specific example of the latter possibility, consider the situation of Figure 4, but with the additional assumptions that air is mixed from the convective region to the subsiding region at an invariable rate, that the temperature is held fixed, and that convective region is saturated with respect to water. Then, if the intensity of the Hadley circulation is increased, evaporation goes up, as does precipitation in the convective region. However, the convective region doesn't get any moister since it is already saturated, while the increased subsidence dries the subtropics. More generally, recall that the "radiatively active" water vapor aloft is a small fraction of the total atmospheric vapor, so that vapor *supply* is not in any event a limiting factor in the water vapor feedback. Only a tiny part of the water evaporated into the atmosphere needs to be diverted into the dry subsiding regions in order to create a large radiative feedback.

5. DISCUSSION

The ultimate trigger for millennial scale climate variations has not yet been identified, but subtropical water vapor changes can be involved in a number of ways. Always lurking in the background is what might be called the "conventional" water vapor feedback, in which a cooler atmosphere contains less moisture in proportion to the reduction in saturation vapor pressure. We have argued that this kind of feedback should apply also to the subtropics, even though the water vapor source is remote, in the convective region, rather than local. This water vapor feedback amplifies the cooling or warming tendencies from any source whatsoever. With regard to the tropical cooling of the Last Glacial Maximum, it would appear that no other form of water vapor feedback is called for; a reduction in CO_2, combined with a modest and entirely plausible increase in oceanic heat export from the tropics appears nearly sufficient to account for the data, when amplified by the conventional water vapor feedback.

Additional coolings or warmings on the order of 2K can be obtained if the subtropical water vapor content is halved or doubled, as compared to the predictions of the "conventional" feedback. Any agent that increases the subtropical subsidence rate will dry the atmosphere beyond the conventional water vapor feedback. An increase in subsidence could be mediated by an increase in sensible heat transported to the extratropics by atmospheric eddies, or an ocean surface cooling induced by enhanced ocean heat transports. Reorganization of convection can also increase the subsidence, as discussed by [Lindzen and Hou (1988); Hou and Lindzen (1992)]. This seems like an especially fruitful possibility to pursue, given that such reorganizations could perhaps be driven by precessional insolation changes indigenous to the tropics. Aside from changes in subsidence, a decrease in large scale mid-tropospheric lateral mixing of moisture could dry the subtropics. The mixing could decrease in response to intensified subtropical jets, a poleward retreat of the extratropical storm tracks, a change in the Northern Hemisphere planetary wave configuration, or a shut-off of some as yet unidentified eddy producing instability indigenous to the tropics. In the domain of ocean-atmosphere collective phenomena, it is also possible that subtle changes in tropical or extratropical conditions could lead to a contraction or expansion of the Pacific warm pool region, with associated changes in low latitude humidity content. To date, extratropical triggers (such as shut off of North Atlantic Deep Water formation) do not seem able to create enough water vapor change in the tropics to produce a global response in climate simulations, but the possibility remains that this failure could be due to inadequate resolution of tropical atmospheric or oceanic dynamics (keeping in mind also the possibility that improvements in dynamic sea ice treatment may enhance the extratropical response itself).

The preceding conclusions regarding water vapor and millennial scale variability are unsatisfyingly vague, and a further examination of the possibilities must rely heavily on general circulation models. There are a few lessons for modelers to be found in this essay. First, a good resolution of tropical dynamics, with regard to transient eddies, jets and subsidence, is essential if one is to accurately capture the subtropical water vapor feedback. The low resolutions employed to date in models of the LGM and subsequent climate variations is a source of concern, and much more will be learned about the subtropical water vapor as models with resolution better than the R15 truncation (about 4 degrees resolution) employed by [Manabe and Broccoli (1985)] become common. Another lesson is that there is a lot going on in the tropical boundary layer that ought to concern us. Processes here govern the boundary layer humidity, surface radiative cooling, evaporation, and generation of low clouds. These things all have a strong effect on the way the tropical circulation and water vapor content change. Diagnostics addressing the issues raised in Section 3 and Section 4 of this essay would be valuable in comparing the way water vapor feedbacks operate in the various simulations. Such diagnostics might include subsidence rate, mid-tropospheric lateral mixing, sensible and latent heat exchange with the midlatitudes, and both boundary layer and mid-tropospheric water vapor mixing ratio. In reporting water vapor changes, it is important to diagnose the upper level water vapor content separately, as these small quantities of water have little impact on column-integrated water content, but nonetheless have a strong radiative impact.

There are many ways tropospheric water vapor could have changed in the course of millennial scale climate variations. A proxy record of paleo-humidity would be invaluable in narrowing down the possibilities. Certainly, boundary layer relative humidity affects the levels of $\delta D$and $\delta^{18}O$ in the boundary layer water vapor, and there may be some way of exploiting this effect in the Tropics. However, what is most of interest for the tropical radiation budget is the free-atmosphere humidity. [Broecker (1997)] argued that tropical glacier $\delta^{18}O$ was indeed such a paleo-hygrometer, but [Pierrehumbert (1999)] pointed out why the isotope data was unlikely to contain much information about the ambient free-troposphere humidity. In [Pierrehumbert (1999)] it was argued that the $\delta^{18}O$ instead is telling us something about the degree to which rainfall over tropical continents (specifically over the Amazon Basin) is lost to runoff. The results indicate that a greater proportion of rainfall was lost to runoff during the Last Glacial Maximum than is lost at present, and argue for a reduction in forest cover during the LGM. Owing to lack of snowline data for later times, it is not known whether similar changes in hydrology are implied in concert with millennial scale variability. The issue is interesting, but its resolution is unlikely to shed much light on the behavior of free-tropospheric water vapor. For the forseeable future, the chief tools one can bring to bear on the problem are likely to be simulation, theory, and development of analogies with presently observable climate variations.

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R. T. Pierrehumbert, Department of Geophysical Sciences, University of Chicago, Chicago, IL 60637. (e-mail: rtp1@geosci.uchicago.edu)

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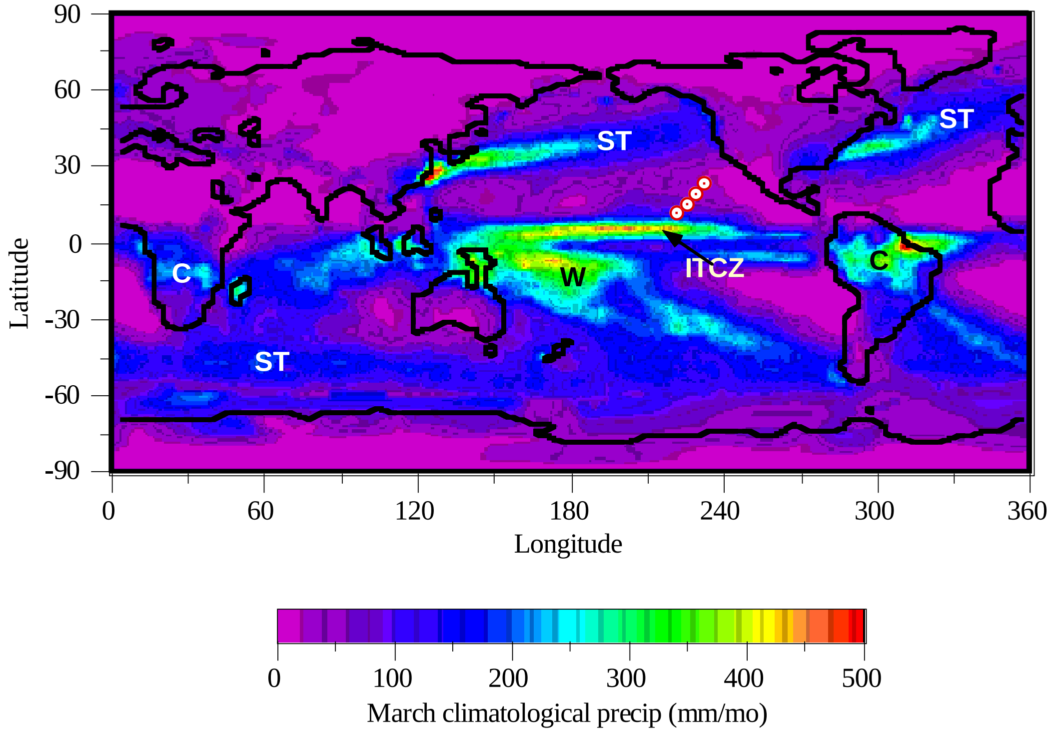

Plate 1. Blended Legates/MSU precipitation climatology for March. Red circles show positions of sondes discussed in the text. The principal convective regions are denoted as follows: ITCZ = Inter Tropical Convergence zone. W = Pacific Warm Pool. C = Tropical Continental. ST = Midlatitude Storm Track.

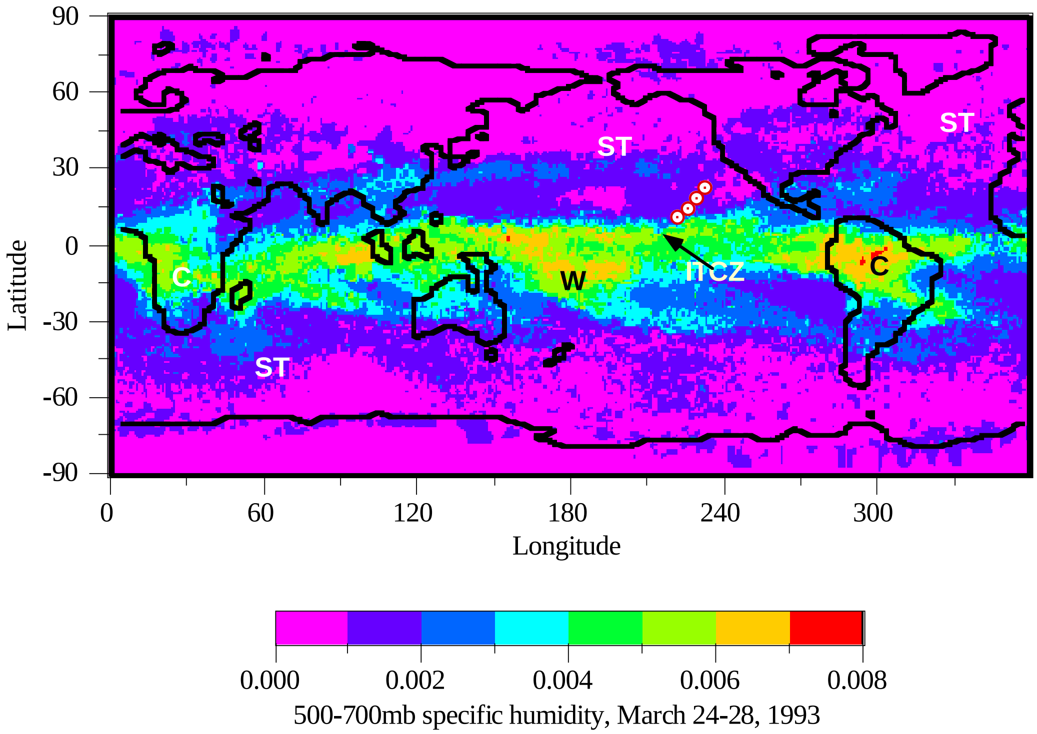

Plate 2. 500-700 mb mean specific humidity from TOVS satellite data, for the mean of the period March 24-29, 1993. Annotations as for Figure 1.

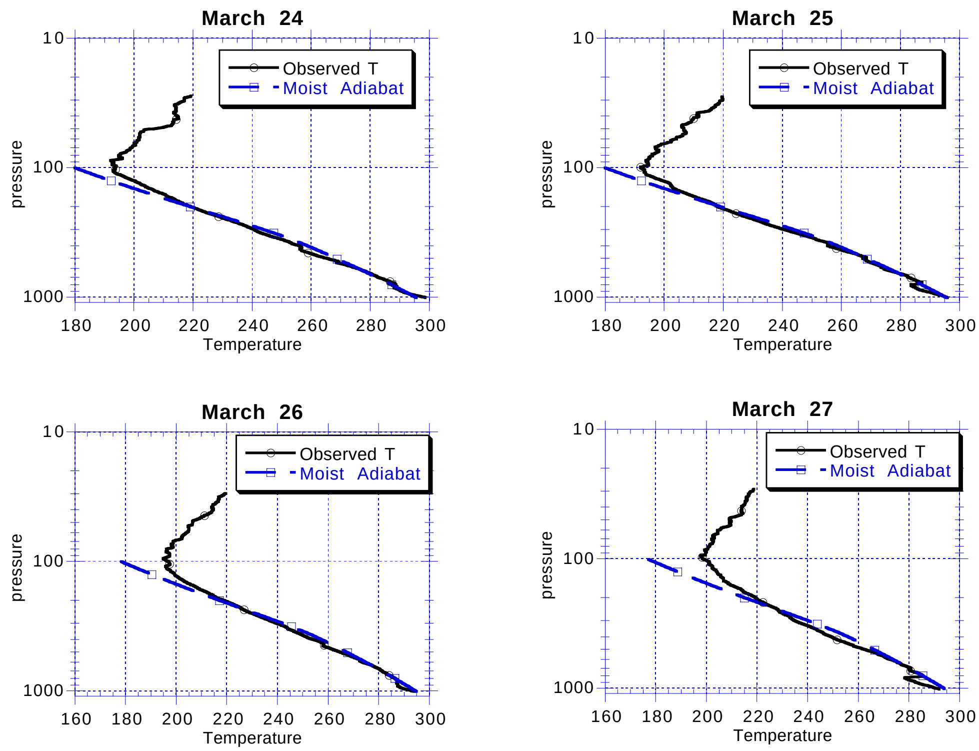

Figure 1. Temperature profiles of the four subtropical CEPEX soundings, compared to the moist adiabat.

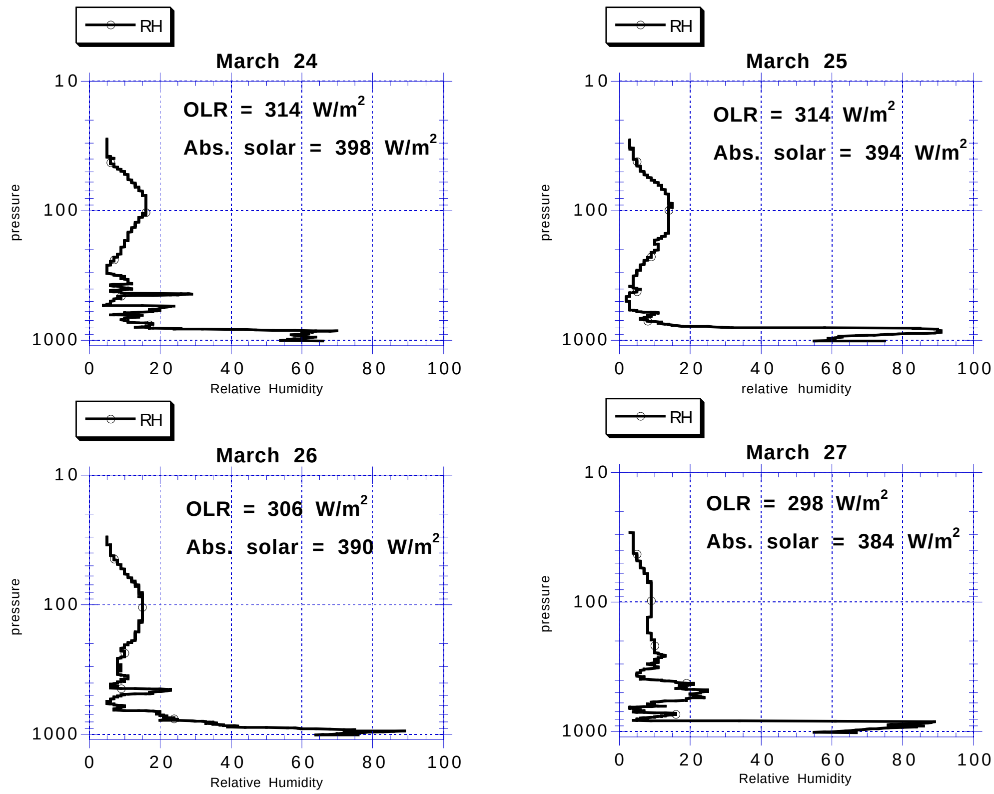

Figure 2. Relative humidity profiles of the four subtropical CEPEX soundings. The clear-sky local absorbed solar radiation and outgoing longwave radiation (OLR) were computed for each sounding using a radiative transfer model, and are indicated on the graphs.

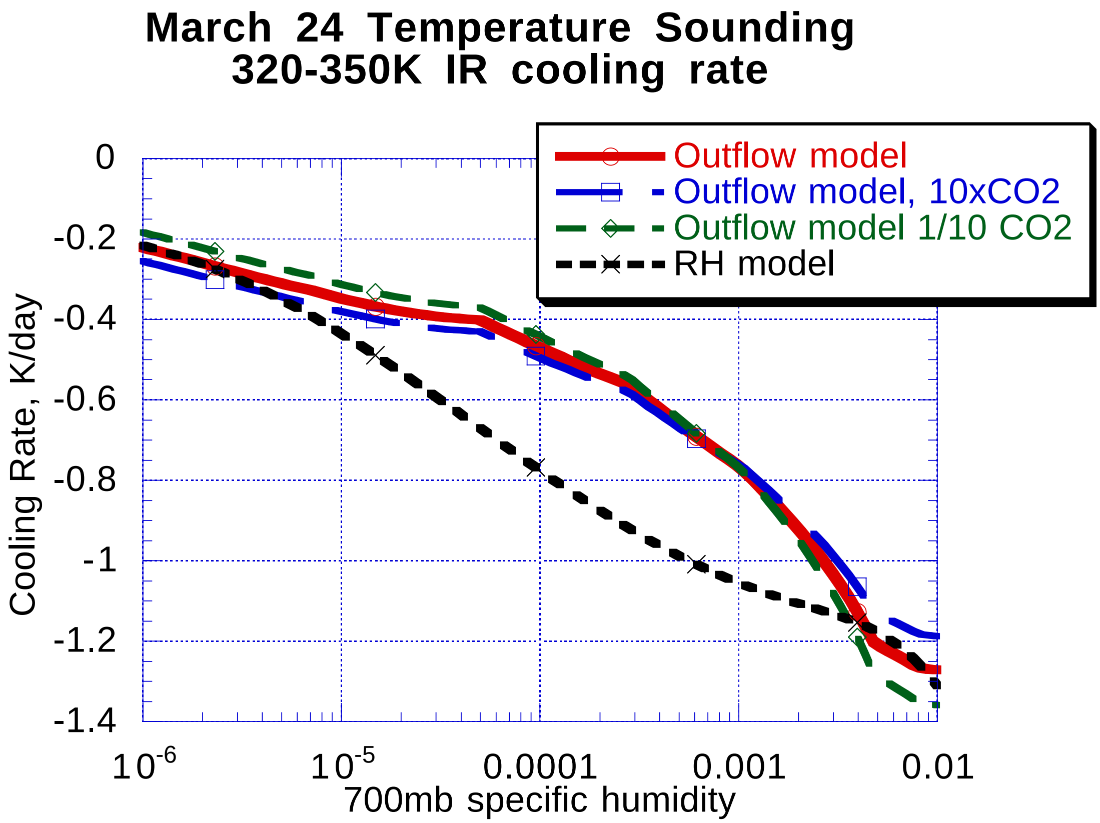

Figure 3. Upper tropospheric infrared radiative cooling for the March 24 temperature sounding, as a function of humidity. Results are shown for the RH model and the outflow model. For the outflow model, the cooling is computed also with 10 times present CO_2 and 1/10 present CO_2. The cooling is averaged over the isentropic layer 320K-350K (about 700-100mb).

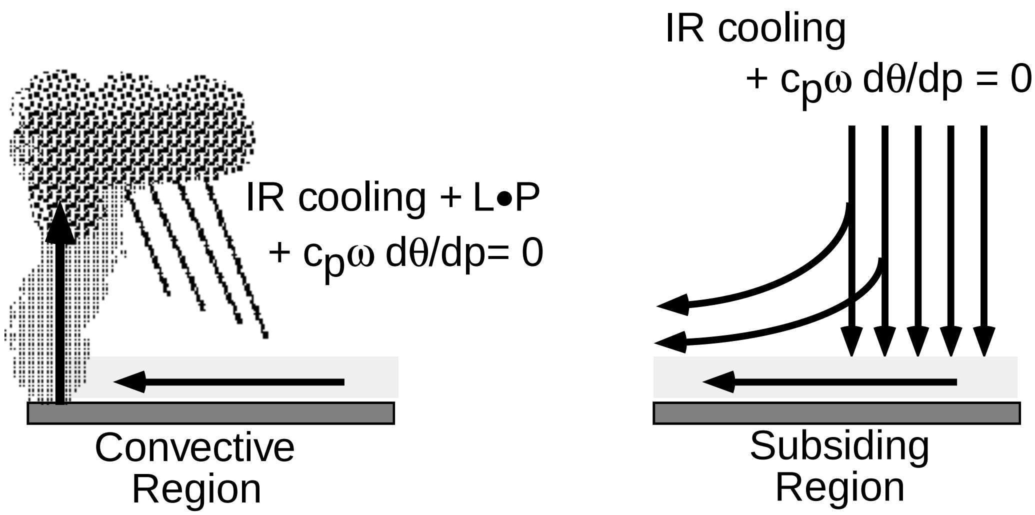

Figure 4. Schematic of radiative/dynamic balance in the convective vs. nonconvective region of the Tropics. In this figure L is the latent heat of condensation, P is the precipitation rate, ω is the pressure velocity and θ is the dry potential temperature.

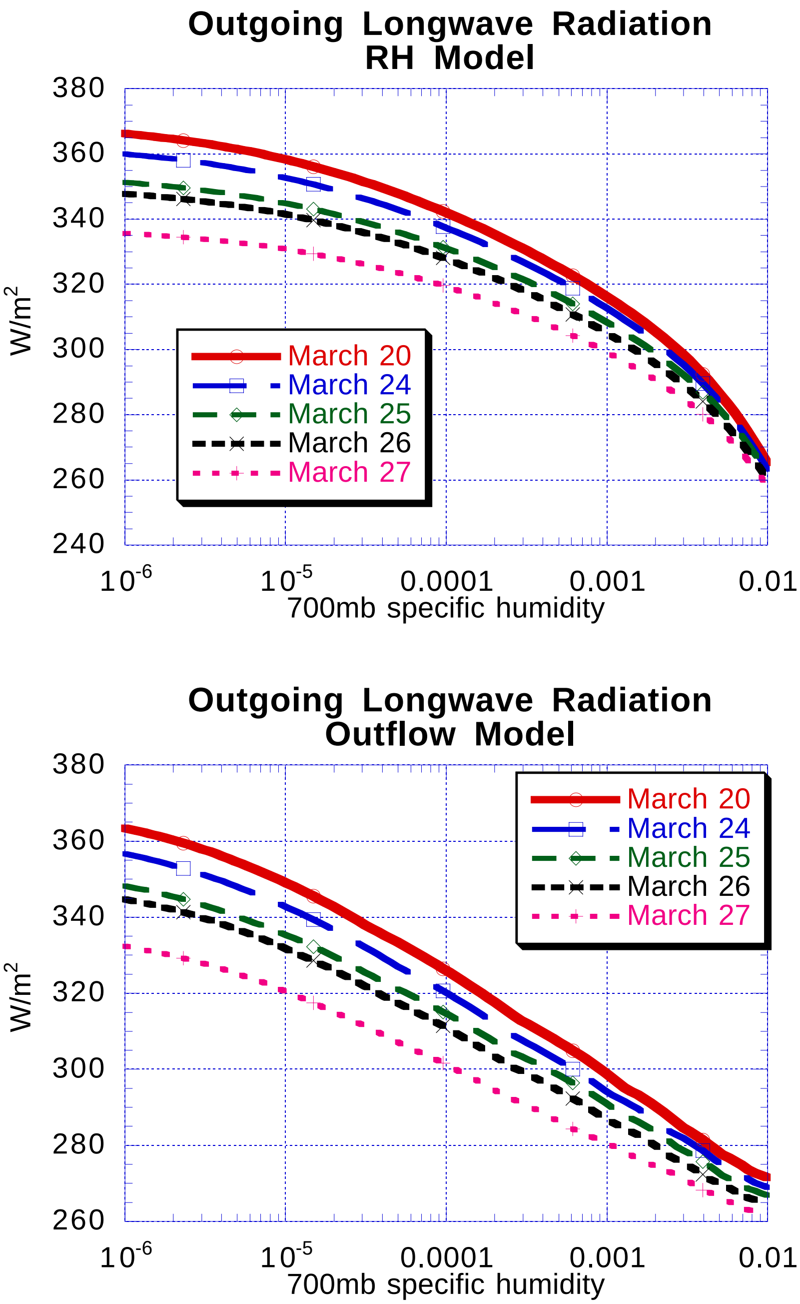

Figure 5. Outgoing longwave radiation as a function of 700mb specific humidity for 5 tropical temperature sounding. Results are shown for the RH model (a) and the outflow model (b).

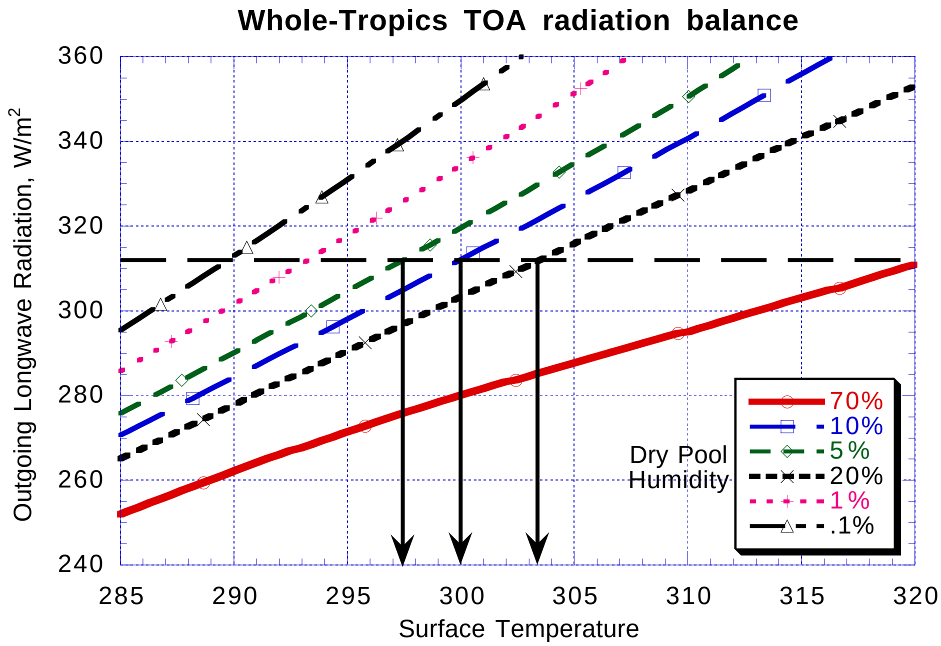

Figure 6. Outgoing longwave radiation averaged over the whole tropics, as a function of surface temperature. Curves are computed for various dry-pool humidities. The vertical profile of humidity was taken according to the RH model. See text for details of the calculation.

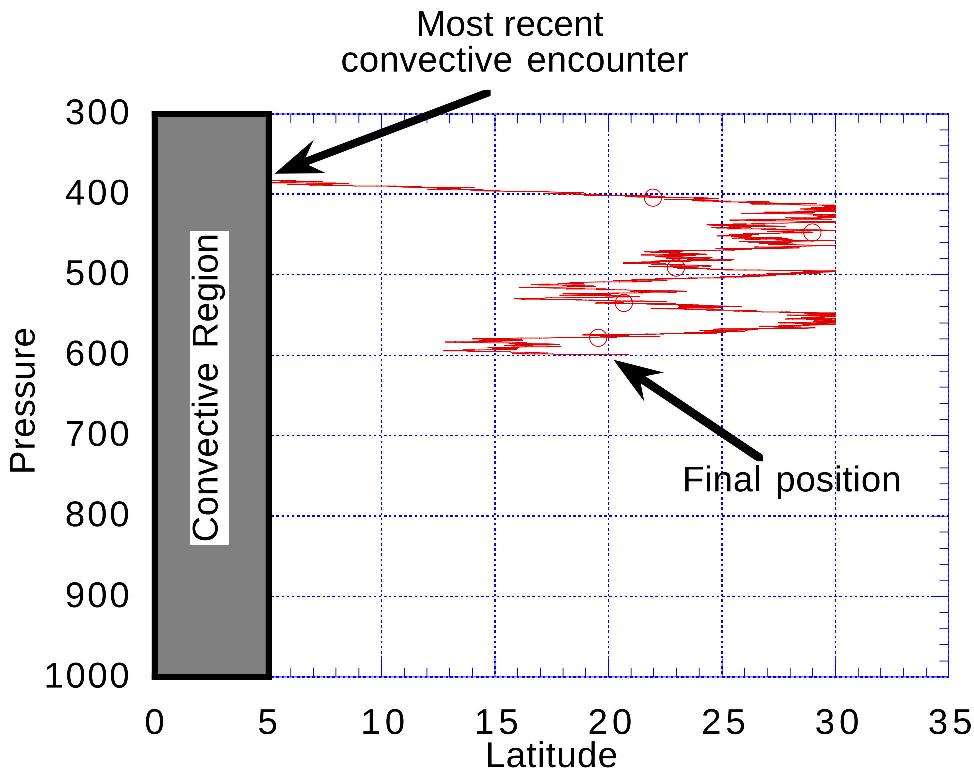

Figure 7. A typical back-trajectory in the advection/subsidence model, showing how p_{min} is determined by the time taken for air to encounter the convective region

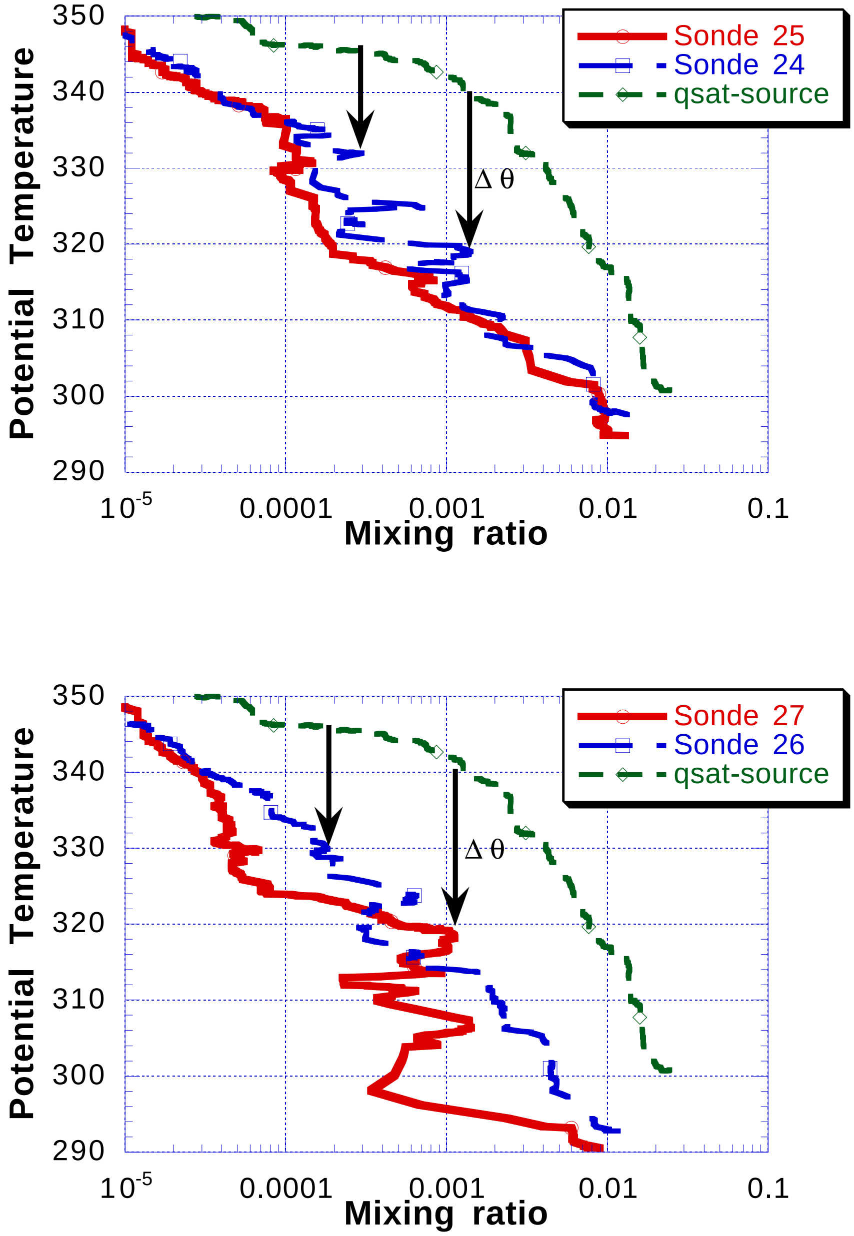

Figure 8. Specific humidity profiles for the four CEPEX subtropical soundings plotted as a function of dry potential temperature. The "source" region sounding is the saturation specific humidity for a sounding taken on the Equator.

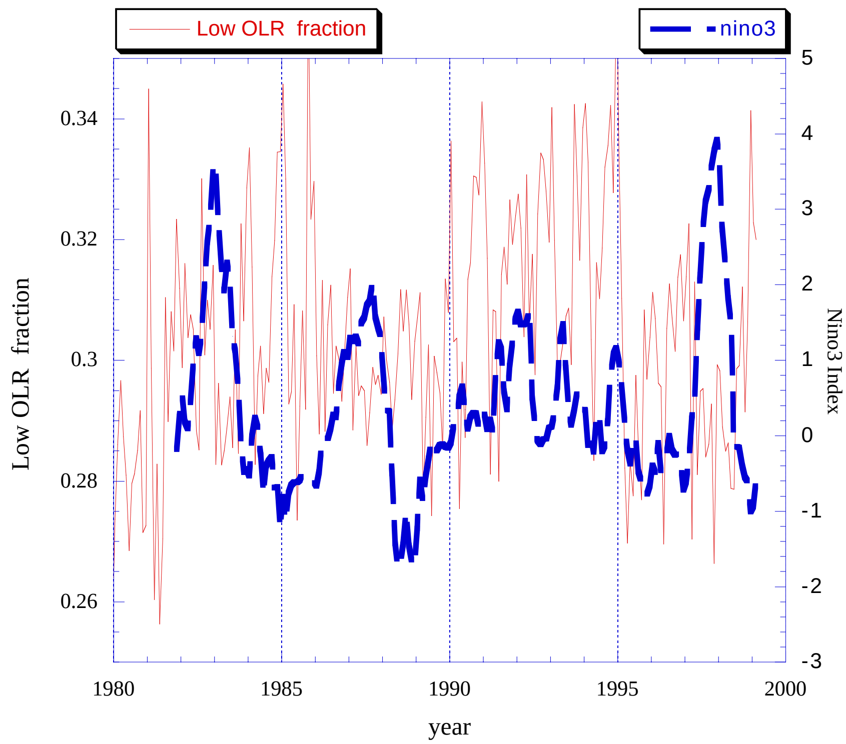

Figure 9. Time series of the Niño3 index, and of a measure of the fractional area covered by the tropical convective region. A high Niño3 index denotes a strong El Niño pattern. The tropical convective index is formed by counting the proportion of the tropical area (25N-25S) for which the monthly-mean OLR is less than 240 W/m^2.