A simple explanation for the increase in relative humidity between 11 and 14 km in the tropics

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We use a very simple model with no adjustable parameters to simulate the increase in relative humidity between 11 and 14 km in the tropics. The rate of increase in relative humidity, in this interval, appears to be largely determined by the shape of the convective detrainment profile, the temperature profile, and the level of zero radiative heating. It appears to be unaffected by evaporative moistening or the complex three-dimensional structure of convective systems. We also show that the rapid increase in the fraction of supersaturated air parcels above 150 mb (~14.5 km) is associated with a transition from large scale descent to ascent.

INDEX TERMS: 0322 Atmospheric Composition and Structure: Constituent sources and sinks; 0365 Atmospheric Composition and Structure: Troposphere—composition and chemistry; 0368 Atmospheric Composition and Structure: Troposphere—constituent transport and chemistry;

KEYWORDS: Water vapor, tropical convection, relative humidity, upper troposphere


1. Introduction

The mean tropical relative humidity profile is shaped like a “C”, with large relative humidities in the boundary layer, a dry mid-troposphere, and large values again near the tropical tropopause [Newell et al., 1997; Jensen et al., 1999; Vomel et al., 2002]. This profile is probably determined by some combination of subsidence drying, export of water vapor from the tropics to the extratropics [Pierrehumbert, 1998; Zhu et al., 2000], moistening by convective detrainment and mixing, and the evaporation of water vapor from falling ice and water [Sun and Lindzen, 1993]. The relative magnitudes of these forcings are still, however, quite uncertain. The weaknesses in our current understanding of the water vapor budget limit our ability to predict how the earth’s climate will respond to the increasing concentrations of carbon dioxide, since, although surface temperatures might be expected to continue to increase in response to higher levels of carbon dioxide, the magnitude of this increase will be strongly affected by the response of water vapor to these higher levels [Manabe and Wetherald, 1967].

Figure 1 shows a compilation of relative humidity with respect to ice (RHI) measurements from three high altitude aircraft campaigns: the 1987 Stratospheric Tropospheric Exchange Project (STEP) [Kelly et al., 1993], the 1993 Central Equatorial Pacific Experiment (CEPEX) [Weinstock et al., 1995], and the 1994 Airborne Southern Hemisphere Ozone Experiment/Measurements for Assessing the Effects of Stratospheric Aircraft (ASHOE/MAESA) [Tuck et al., 1997]. Each relative humidity measurement represents a 10 second average, and comes from within the 20°S–20°N latitudinal band. Most of the STEP measurements were taken in the vicinity of Darwin in northern Australia, while most of the ASHOE/MAESA and CEPEX measurements occurred in the vicinity of Fiji. The water vapor measurements from the STEP and ASHOE/MAESA campaigns were obtained using the NOAA Aeronomy Laboratory Lyman α hygrometer, while those from the ER-2 during the CEPEX campaign were taken using the Harvard Lyman α hygrometer [Weinstock et al., 1994]. A discussion of these instruments is given in the SPARC Assessment of Upper Tropospheric and Stratospheric Water Vapor [Kley et al., 2000]. Temperature and pressure measurements from STEP and ASHOE/ MAESA were made by the Micrometeorological Measurement System (MMS), which has a temperature accuracy at these altitudes of ±0.3 K, and a pressure accuracy of ±0.25 mb [Chan et al., 1989]. Temperature and pressure measurements from the ER-2 during CEPEX were made using the NAV recorder, whose associated errors are likely to be larger than those from the MMS.

The curve in Figure 1 with open boxes is an average of the relative humidity measurements in pressure intervals. Relative humidity measurements above 1.6 are occasionally present but have little impact on the average climatology. The mean relative humidity increases from about 0.5 below
This paper addresses the reasons for this increase.

2. Model

Figure 2 shows a simplified moisture budget of the upper tropical troposphere. The mean water vapor mixing ratio is assumed to equal a weighted average over air parcels which have detrained at various heights from deep convective clouds. Air parcels which detrain above the level of zero radiative heating (about 15 km) will be subject to mean ascent and are unlikely to influence the water vapor budget at lower altitudes. It is assumed that the water vapor mixing ratio of an air parcel at detrainment is equal to the saturation water vapor mixing ratio at that temperature, and that this mixing ratio is conserved during subsidence. This ignores possible increases in the water vapor mixing ratio due to evaporation of ice crystals falling from cirrus clouds at higher altitudes. The relative contribution of a given pressure interval to the mean water vapor mixing ratio is proportional to the detrainment rate at that pressure.

\[ \left[ H_2O(p) \right] = \frac{\int_{p}^{p_{O-0}} [H_2O_{sat}(p')] \text{det}(p') dp'}{\int_{p}^{p_{O-0}} \text{det}(p') dp'}, \tag{1} \]

where \( p_{O-0} \) is the pressure at which the clear sky radiative heating is zero, and \( [H_2O_{sat}(p')] \) is the saturation water vapor mixing ratio corresponding to the temperature at pressure \( p' \).

In this model, the mean water vapor mixing ratio in the upper tropical troposphere is purely a function of the mean temperature and detrainment profiles, and the level of zero radiative heating. A detrainment profile which was sharply peaked near \( p_{O-0} \) would give rise to smaller mean water vapor mixing ratios than one in which deep convective outflow tended to occur at a lower altitude where the saturation water vapor mixing ratios at detrainment were larger.

The dashed lines in Figure 3 show two relative humidity profiles generated by the model. The dashed line with the smaller relative humidity has been calculated using (1). The larger of the two dashed curves was also calculated from (1), but with the detrainment temperature increased by 1 K. The reasons for this shift will be discussed later. Between 11 and 14 km, the two curves agree reasonably well with the aircraft climatology (open boxes), and have been converted to pressure using a climatological relationship between height and pressure. Also shown in Figure 3 is a relative humidity climatology compiled from approximately 800 sondes launched at nine different tropical locations as part of the SHADOZ project (Southern Hemisphere Additional Ozonesondes project [Thompson et al., 2002]). Below the 0°C line (~600 mb),
The relative humidity has been calculated with respect to the saturation vapor pressure of water rather than ice.

3. The Detrainment Profile

The detrainment profile used in (1) is shown in Figure 4. It was determined from the tropical mean clear sky mass flux $M_r(p)$ using

$$det(p) = g \frac{dM_r(p)}{dp}$$  \hspace{1cm} (2)

The mass flux was determined from the clear sky heating rate $Q_r(p)$ using [Minschwaner and McElroy, 1992]

$$M_r(p) = \rho \frac{Q_r(p)}{dT/dz + \Gamma_d}$$  \hspace{1cm} (3)

where $g$ is the gravitational acceleration, $\rho$ the density, $dT/dz$ the lapse rate, and $\Gamma_d$ the dry adiabatic lapse rate. The clear sky radiative heating rate $Q_r(p)$ was calculated using a δ-four-stream radiative transfer code [Fu and Liou, 1992], with the input temperature, water vapor, and ozone profiles obtained from a compilation of radiosonde and satellite measurements [Folkins, 2002]. These input profiles were generated for each 5° latitude interval from 20°S to 20°N, and used to calculate heating rate and mass flux profiles within each interval. The mass flux profile, shown in Figure 5 was obtained by averaging over all mass flux profiles from 20°S to 20°N. This mass flux profile was then used in (2) to generate the detrainment profile shown in Figure 4.

The use of (3) to estimate the downward mass flux in the tropics involves three main assumptions. First, it assumes that the area covered by slow radiatively driven subsidence is much larger than the area covered by cloud updrafts. This is probably a good assumption because most upward transport in the tropics occurs within deep convective clouds, which cover about 2.7% of the total area [Rossow and Schiffer, 1999]. Second, it assumes that the downward transport of mass associated with evaporative cooling from falling ice and rain can be ignored. This is a poor assumption in the mid and lower troposphere where the vertical mass fluxes associated with convective downdrafts are substantial, but it is probably a reasonable assumption in the upper troposphere. This is because, although falling ice crystals may evaporate and bring the relative humidity has been calculated with respect to the saturation vapor pressure of water rather than ice.

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by deep convective clouds, the entrainment of mass out of the layer by deep convection, and by the exchange of mass in that layer with higher latitudes. The entrainment of mass from a layer should be small if the mass fluxes associated with upper tropospheric convective downdrafts are small. Estimates of horizontal transport from radiosonde winds suggest that the exchange of mass with higher latitudes can be considered small in the 11 to 14 km interval where the vertical divergence is large [Folkins, 2002].

Although the use of clear sky radiative heating rates to infer tropical mean detrainment rates is subject to a large number of assumptions, and can only be expected to be valid in the upper troposphere, some features of the inferred detrainment profiles are consistent with independent measurements. Lidar measurements from TOGA/COARE indicate that the mean base and top of deep convective anvil cirrus occur near 12 km and 16.85 km respectively [Sassen et al., 2000]. Using a relationship between potential temperature and height, appropriate for this region and season, this height range corresponds to a potential temperature range of 348 K for cloud base to 368 K for cloud top [Folkins, 2002]. This is roughly the range over which the detrainment profile shown in Figure 4 is positive. It was also shown that, when implemented in a one-dimensional model, the detrainment profile shown in Figure 4 could be used to generate an ozone profile similar to observed climatologies [Folkins et al., 2002].

Figure 6 shows a compilation of 10 sec averaged ozone measurements from the STEP and ASHOE/MAESA aircraft campaigns. At marine tropical locations, such as Darwin (STEP) and Fiji (ASHOE/MAESA), ozone mixing ratios in the boundary layer (below 800 mb) are usually less than 20 ppbv. The highest potential temperature at which such low ozone values are found in the upper troposphere is 365 K. This suggests that the 365 K potential temperature surface is near the upper limit of moist undilute convective ascent from the boundary layer. This upper limit is consistent with Figure 4, which shows that the detrainment rate passes through zero near 365 K. One would also expect the rate of undilute convective detrainment to be small above 365 K because air parcels in the boundary layer with \( \theta_e \) larger than 365 K are rarely observed [Folkins, 2002]. In most of the tropics, the 365 K potential temperature surface corresponds to an altitude of about 16 km. It corresponds to a somewhat higher altitude in regions such as the western tropical Pacific, where potential temperatures near the cold point tropopause tend to be displaced upward.

The most remarkable feature of the detrainment profile shown in Figure 4 is the rapid increase near 230 mb (~11 km). A rapid increase in detrainment near this pressure has been previously inferred from an array of radiosonde profiles in the western tropical Pacific [Yanai et al., 1973]. From a thermodynamic point of view, air parcels near the surface become able to participate in deep convection once their Convective Available Potential Energy (CAPE) is positive. In most of the tropics, the threshold pseudoequivalent potential temperature \( \theta_e \) at which the CAPE of an air parcel becomes positive is near 345 K. If deep convective updrafts are able to transport nearly undilute boundary layer air into the upper tropical troposphere (so that the \( \theta_e \) of these air parcels is conserved and they detrain near their level of neutral buoyancy), then one
would expect the onset of deep convective detrainment to occur near the 345 K isentropic surface [Folkins, 2002].

4. Supersaturated Fraction

[13] Figure 1 shows that there is a greater abundance of supersaturated (RHI > 1) air parcels in the tropical troposphere at higher altitudes. In Figure 5, the fraction of supersaturated air parcels has been plotted as a function of pressure. This fraction rapidly increases above 150 mb (~14.5 km) as the level of zero radiative heating is approached from below. Between the level of zero radiative heating and the cold point tropopause, large scale ascent will tend to increase the relative humidity, and, in the absence of ice condensation nuclei, will tend to generate relative humidities greater than one [Jensen et al., 1999]. Below the level of zero radiative heating, large scale descent can be expected to significantly reduce the frequency of supersaturated air parcels.

5. Undulation of Potential Temperature Surfaces

[14] The temperatures used to calculate the relative humidities shown in Figure 1 are given in Figure 7. Superimposed on the STEP, ASHOE/MAESA, and CEPEX temperature measurements is a black line representing a 15°S–15°N annual climatology. Temperatures from STEP and ASHOE/MAESA exhibit a warm bias below 120 mb (~15.3 km) and a cold bias above 120 mb. Temperatures from CEPEX were closer to the tropical mean. The most likely reason for the anomalous temperatures observed during these campaigns is
that they occurred in regions of active deep convection. In the case of STEP and ASHOE/MAESA, a likely contributing factor to the cold bias above 120 mb is that these measurements occurred during Northern Hemisphere winter. The pressure at which these temperature anomalies change sign (~120 mb) is quite close to the pressure at which clear sky radiative heating rates change sign (~125 mb), suggesting that convectively induced temperature anomalies are subject to radiative damping.

The existence of temperature anomalies associated with tropical deep convection introduces ambiguities into the interpretation of relative humidity measurements obtained primarily from convective regions. The existence of a warm bias in convective regions would tend to reduce the relative humidity in these regions with respect to the tropical mean. This might explain why, as shown in Figure 3, the aircraft climatology is smaller than the MLS climatology between 12 and 14 km. On the other hand, this effect would be offset by a tendency for the relative humidity in convective regions to be larger than the tropical mean because air parcels in these regions have detrained more recently from convective clouds.

If convective regions tend to be warmer than the tropical mean, then the use in (1) of a saturation water vapor mixing ratio at a detrainment temperature corresponding to the tropical mean at that pressure may be inappropriate. As a sensitivity test, we increased the detrainment temperature in (1) by 1 K. The resulting relative humidity profile is shown as the larger of the two dashed lines in Figure 3. Increasing the detrainment temperature by a constant shifts the modeled relative humidity profile weighted more strongly to a higher altitude would produce a relative humidity profile that had lower values at 11 km, and that increased with height more rapidly than observed.

Also identified in Figure 3 is the Tropical Tropopause Layer (TTL) [Highwood and Hoskins, 1998; Folkins et al., 1999]. This layer can be defined as the interval over which the mean temperature profile is controlled by both the Hadley and Brewer Dobson Circulations [Folkins, 2002]. The presence of large scale ascent in most of the TTL renders the model inapplicable in this region.

Figure 3 shows that the modeled relative humidity profiles agree quite well with the observed climatologies between 11 and 14 km. This is coincident with the height range over which the detrainment rate appears to be controlled by the probability distribution function of $\theta_0$ in the convective boundary layer [Folkins, 2002]. It has been identified as the scaling layer in Figure 3. In the model, relative humidity increases with height in the scaling layer because the amount of post detrainment subsidence becomes progressively less as the level of zero radiative heating is approached from below. For example, the mean relative humidity at 11 km arises from an average over very moist air parcels which have recently detrained near 11 km and subsided very little, and very dry air parcels which have detrained near the level of zero radiative heating (~15.3 km) and subsided by more than 4 km. For air parcels at 15 km, however, the mean detrainment height can only be slightly larger than 15 km, so one would expect to observe a mean relative humidity only slightly less than 1. The good agreement between the model and the measurements in the rate at which relative humidity increases with height in the scaling layer provides some support that the detrainment profile shown in Figure 4 is realistic. For example, a detrainment profile weighted more strongly to a higher altitude would produce a relative humidity profile that had lower values at 11 km, and that increased with height more rapidly than observed.

The effects occur in reverse above 120 mb.

6. Discussion

As shown in Figure 3, the modeled relative humidity profiles agree quite well with the observed climatologies between 11 and 14 km. This is coincident with the height range over which the detrainment rate appears to be controlled by the probability distribution function of $\theta_0$ in the convective boundary layer [Folkins, 2002]. It has been identified as the scaling layer in Figure 3. In the model, relative humidity increases with height in the scaling layer because the amount of post detrainment subsidence becomes progressively less as the level of zero radiative heating is approached from below. For example, the mean relative humidity at 11 km arises from an average over very moist air parcels which have recently detrained near 11 km and subsided very little, and very dry air parcels which have detrained near the level of zero radiative heating (~15.3 km) and subsided by more than 4 km. For air parcels at 15 km, however, the mean detrainment height can only be slightly larger than 15 km, so one would expect to observe a mean relative humidity only slightly less than 1. The good agreement between the model and the measurements in the rate at which relative humidity increases with height in the scaling layer provides some support that the detrainment profile shown in Figure 4 is realistic. For example, a detrainment profile weighted more strongly to a higher altitude would produce a relative humidity profile that had lower values at 11 km, and that increased with height more rapidly than observed.

Figure 3 shows that the modeled relative humidity is much lower than the observed climatologies below 11 km, so that the model is clearly inapplicable in this region also. Below 11 km, there is virtually no detrainment in the model, so that the modelled vapor mixing ratio is essentially equal to the mass weighted average of air parcels that have detrained in the scaling layer (~100 ppmv). This generates unrealistically low relative humidities in most of the tropical troposphere. This is presumably due to the absence from the model of processes such as shallow convection, evaporation of falling water and ice, and mixing along the sides of deep convective updrafts, which prevent relative humidities from approaching near zero values in the mid-troposphere. It should be noted, however, that the absence of these processes does not produce a relative humidity profile corresponding to the saturation water vapor mixing ratio of the tropical tropopause, as is sometimes thought. This would only occur if convective detrainment could be described as a $\delta$ function spike at the tropical tropopause.

It is clear from (1) that the multiplication of the detrainment profile by a constant, with no change in tropical temperatures or the level of zero radiative heating, has no effect on the relative humidity profile generated by the model.
This suggests that a change in the intensity of the Hadley circulation should, by itself, have little effect on the mean relative humidity profile of the upper tropical troposphere.

[23] There have been several recent demonstrations that the mean relative humidity field in various parts of the tropics can be well simulated by large-scale advection, with the restriction that water vapor be at or below saturation [Sherwood, 1996; Salathe and Hartmann, 1997; Pierrehumbert and Roca, 1998; Gettelman et al., 2000]. In particular, it has been shown, on the 215 and 146 mb surfaces, that the MLS relative humidity measurements can be reproduced by a three-dimensional model using realistic large scale winds, and a relative humidity cutoff of 1, but with no attempt to incorporate cloud microphysical processes [Dessler and Sherwood, 2000]. The success of this approach, in the language of this paper, may have arisen from the fact that both pressure surfaces lie within the scaling layer.

[23] Within a convective system, one would expect the evaporation of falling ice and water to increase the relative humidity of unsaturated air below the cirrus anvils, but not in the detraining air itself, since this air should be already close to saturation. The mean cloud base of deep convective cirrus anvils, at least in the western Pacific, is near 12 km [Sassen et al., 2000]. This may help account for the absence of a need for an evaporative moisture source between 11 km and 14 km in the tropics.

7. Summary

[24] The water vapor budget of the tropical troposphere is very complicated, with fundamental quantities such as the mean precipitation efficiency still poorly constrained by measurements. We have shown, however, that for some portion of the upper tropical troposphere - the interval from 11 to 14 km - the mean relative humidity profile is determined by the deep convective detrainment profile, the temperature profile, and the height of the level of zero radiative heating.

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References


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