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Tropical convective outflow and near surface equivalent potential temperatures

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Abstract. We use clear sky heating rates to show that convective outflow in the tropics decreases rapidly with height between the 350 K and 360 K potential temperature surfaces (or between roughly 13 and 15 km). There is also a rapid fall-off in the pseudoequivalent potential temperature probability distribution of near surface air parcels between 350 K and 360 K. This suggests that the vertical variation of convective outflow in the upper tropical troposphere is to a large degree determined by the distribution of sub cloud layer entropy.

Subcloud layer entropy plays an important role in determining the depth of convective mixing in the tropics [*Emanuel*, 1994]. Figure 1 shows two probability distributions of "near surface" pseudoequivalent potential temperature θ_e in the tropics. θ_e is a measure of entropy which includes the potential contribution arising from water vapor condensation. Its virtue in this context is that the θ_e of an air parcel is approximately conserved during upward transport within clouds provided there is no mixing. The cumulative probability distribution $N(\theta_e)$ shown in Figure 1 gives the fraction of air parcels below 1 km whose pseudoequivalent potential temperature is larger than a given θ_e. Approximately 7 % of the air parcels below 1 km in the tropics have a θ_e larger than 355 K. The curve $dN(\theta_e)/d\theta$ is the relative probability distribution of near surface θ_e obtained by counting the number of θ_e measurements in 1 K bins. It has been normalized so that its largest value is equal to 1. There is a rapid (roughly exponential) decrease in the occurrence of air parcels with θ_e larger than 355 K.

The distribution shown in Figure 1 is intended to represent an annual tropical average. It was compiled from temperature and humidity profiles from the following missions: the Pacific Exploratory Mission (PEM) - Tropics A and B, the Indian Ocean Experiment (INDOEX), and the Southern Hemisphere Additional Ozonesondes (SHADOZ) initiative. For simplicity, we will collectively refer to this dataset as the SHADOZ archive. It involved a total of 673 profiles - 93 from Ascension Island (8 °S), 4 from Christmas Island (2 °N), 77 from San Cristobal (1 °S), 114 from Fiji (18 °S), 23 from Watukosek Java (7.5 °S), 53 from Kaashidhoo (5 °N), 94 from Nairobi (1.3 °S), 40 from Natal (5.4 °S), and 175 from Samoa (14.2 °S). The average latitude, weighted by the number of profiles, was 8.7 °S. There was an approximately equal number of sondes taken in each season. Since the sondes were taken at varying vertical resolution, a $N(\theta_e)$ was first calculated for each site, and then entered into the overall average in a weighted manner depending on the number of sondes at that site. Pseudoequivalent potential temperatures were calculated using the Bolton parameterization [*Bolton*, 1980].

The two main mechanisms which drive downward descent across isentropic surfaces in the tropics are radiative and evaporative cooling. The radiatively driven mass flux $M_r(\theta)$ can be approximated as

$$M_r(\theta) = \rho(\theta)w_r(\theta) = \rho(\theta)Q(\theta)/(dT(\theta)/dz + \Gamma_d). \qquad (1)$$

Here $dT(\theta)/dz$ is the derivative of temperature with respect to height, Γ_d the dry adiabatic lapse rate, $\rho(\theta)$ density, $w_r(\theta)$ the downward radiative velocity, and $Q(\theta)$ the clear sky radiative cooling rate. This expression is valid in clear-sky regions over long timescales where the heights of potential temperature surfaces can be regarded as fixed.

The left hand side of Figure 2 shows the vertical variation of $M_r(\theta)$ in the upper tropical troposphere obtained from (1). The open circles represent the mass flux $M_r(\theta)$ while the solid circles represent $N(\theta_e)$, with in this case, the vertical axis understood to be θ_e rather than θ. The horizontal axes of $M_r(\theta)$ and $N(\theta_e)$ have been adjusted so that the two plots roughly coincide at $\theta = 346$ K. This has been done to emphasize that $M_r(\theta)$ and $N(\theta_e)$ are roughly proportional to one another for θ (or θ_e) larger than 345 K.

The simplest way to explain the scaling between $M_r(\theta)$ and $N(\theta_e)$ shown in Figure 2 is to assume (1) that near surface air parcels have a roughly equal probability of being subjected to deep convection, (2) that air parcels detrain from convective systems near their Level of Neutral Buoyancy (LNB), (3) that most of the downward descent in the tropics is driven by radiative cooling, and (4) that the radiative effect of clouds on the vertical mass flux is small. Although none of these assumptions can be expected to be correct in any exact sense, they may be sufficiently valid above 345 K to account for the scaling between $M_r(\theta)$ and $N(\theta_e)$ in this region.

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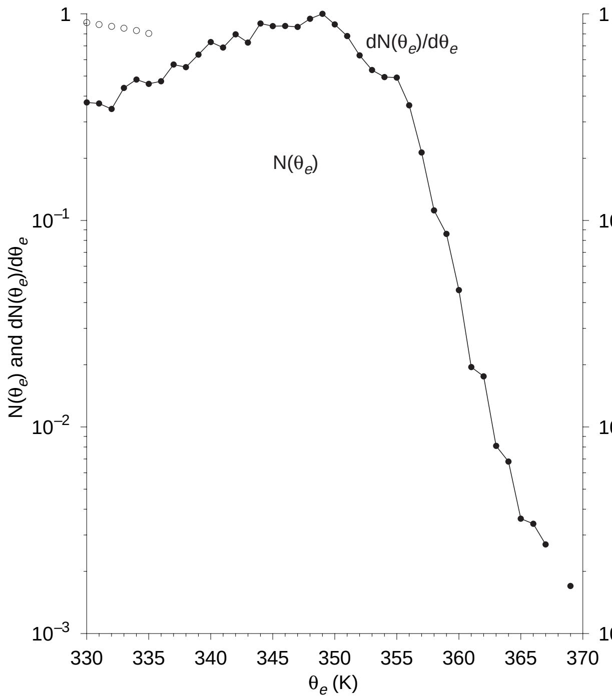

Figure 1. Two probability distributions of near surface (below 1 km) θ_e values in the tropics. $N(\theta_e)$ is the cumulative probability distribution representing the fraction of air parcels whose θ_e is larger than a given value. $dN(\theta_e)/d\theta$ is the relative probability distribution obtained by counting the number of θ_e measurements within a given θ_e interval, in this case 1 K. The peak of this distribution has been arbitrarily normalized to 1. Both distributions were obtained from the SHADOZ archive.

The right hand side of Figure 2 shows that the vertical divergence of the radiatively driven mass flux, $dM_r(\theta)/dz$, roughly scales with the relative near surface θ_e probability distribution $dN(\theta_e)/d\theta_e$. This follows from the scaling between $M_r(\theta)$ and $N(\theta_e)$ and is shown for illustrative purposes. The divergence of $M_r(\theta)$ can be regarded as approximately equal to that part of the convective outflow which is balanced by radiative cooling.

A three-dimensional modeling simulation gives the net mass flux at 10 km as approximately 0.004 $\mathrm{kgm^{-2}s^{-1}}$, or 345 $\mathrm{kgm^{-2}day^{-1}}$ [*Tompkins and Craig*, 1998]. Net mass flux in this case refers to the residual between the mass fluxes from convective updrafts and downdrafts, and should approximately equal M_r at 10 km, which from Figure 2 is about 350 $\mathrm{kgm^{-2}day^{-1}}$.

The clear-sky heating rates used in (1) were calculated using a radiative transfer model based on the δ-four-stream method [*Fu and Liou*, 1992]. The annual tropical climatologies of temperature, water vapor, ozone, and pressure at 0.25 km resolution used in the model was obtained from the SHADOZ archive. Because of uncertainties in the accuracy of water vapor measurements above 12 km, we imposed a constant 3.8 ppmv water vapor mixing ratio above 17 km [*Dessler*, 1998], and used spline interpolation to determine the relative humidity between 12 and 17 km.

The degree of scaling between $M_r(\theta)$ and $N(\theta_e)$ is extremely sensitive to the shape of the input temperature profile used to calculate the radiative heating rate. We used the Forecast Systems Laboratory/National Climatic Data Center (FSL/NCDC) Radiosonde Archive to construct tropical temperature climatologies defined between various latitudinal ranges. The SHADOZ temperature profile most nearly corresponds to an average over a latitudinal range of 25 °S to 25 °N. Temperature climatologies constructed from a more

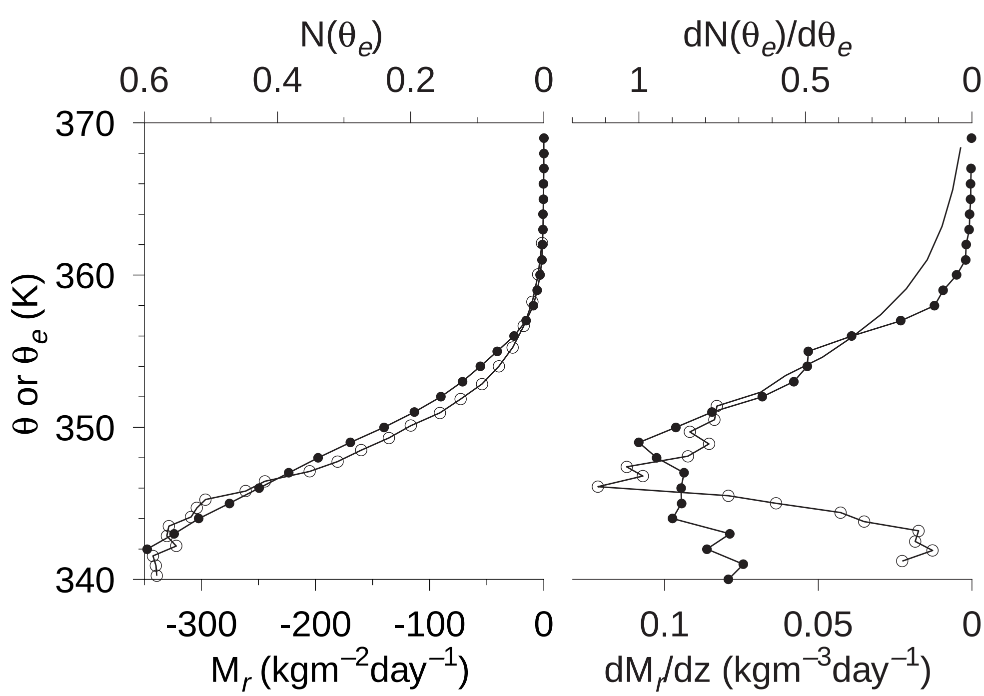

Figure 2. (left) The curve with open circles is the clear sky vertical mass flux $M_r(\theta)$ as obtained from (1) in the text. The curve with solid circles is $N(\theta_e)$ from Figure 1, but with the axes interchanged. (right) The curve with open circles is $dM_r(\theta)/dz$. The curve with solid circles is $dN(\theta_e)/d\theta$, also from Figure 1 with the axes interchanged.

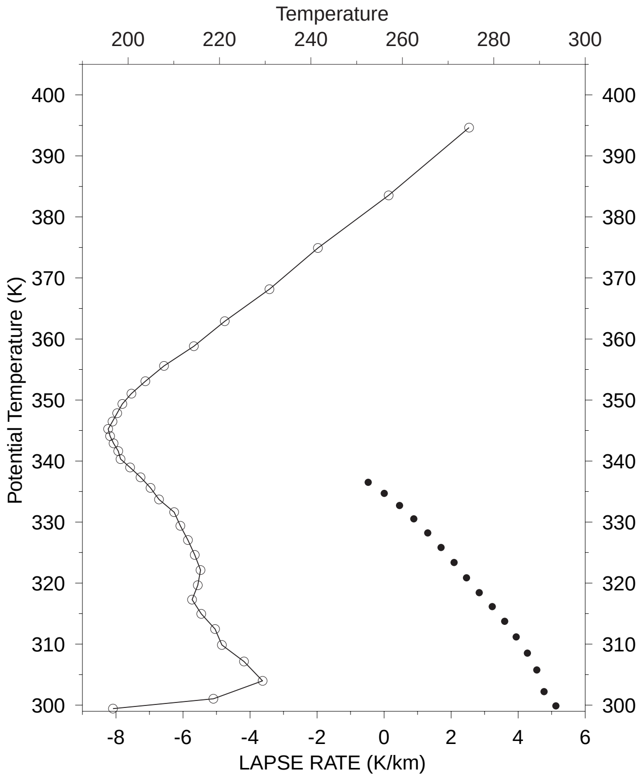

Figure 3. Temperature and lapse rate (here dT/dz rather than -dT/dz) versus potential temperature, from the SHADOZ archive. The lapse rate tropopause is usually defined as the altitude at which temperature decreases less rapidly than 2 K/km.

restricted latitudinal range closer to the equator are somewhat colder than the SHADOZ climatology between 350 and 380 K. This tends to make $M_r(\theta)$ decrease more steeply with θ above 345 K and strongly degrade the scaling between $M_r(\theta)$ and $N(\theta_e)$. Since this scaling is a statistical relationship which can be expected to be valid only over sufficiently long times and large distances, it perhaps makes sense that the input temperature profile should be an average over the full width of the Hadley circulation.

The SHADOZ temperature and lapse rate profiles used in the radiative transfer model are plotted versus θ in Figure 3. The onset of scaling between $M_r(\theta)$ and $N(\theta)$, the peak of the near surface θ_e relative probability distribution, and the onset of an increase in lapse rate all occur between 345 and 350 K. By 370 K, the convective outflow (as diagnosed from the radiative mass flux) and the near surface θ_e relative probability distribution have essentially diminished to zero. This is below both the cold point (383 K) and lapse rate (375 K) definitions of the tropical tropopause, and is consistent with previous work based on GCM's [*Thuburn and Craig*, 1997], cloud top temperatures [*Highwood and Hoskins*, 1998], and ozone and lapse rate variations [*Folkins et al.*, 1999], that the traditional definitions of the tropical tropopause are not good indicators of the mean upper limit of convective mixing in the tropics.

Figure 4 shows tropical monthly climatologies of potential temperature versus height between 15 °S and 15 °N constructed from the FSL/NCDC Radiosonde Archive. Spline interpolation was used to find temperatures between the standard pressure levels. As can be seen from Figure 4, temperatures in the tropical lower stratosphere are lowest during Northern Hemisphere winter when the adiabatic cooling induced by upward ascent is strongest [*Yulaeva et al.*, 1994]. This seasonal cycle extends down to roughly 355 K (14.5 km), the approximate altitude at which the convective outflow shown in Figure 2 has diminished to half its maximum value.

Airborne lidar measurements in January and February during the 1993 Tropical Ocean and Global Atmosphere (TOGA) Coupled Ocean-Atmosphere Regional Experiment (COARE) frequently detected cloud tops at and sometimes above the average tropopause height of 16.85 km [*Sassen et al.*, 2000]. This appears to be in contradiction with the mean θ_e probability distributions shown in Figure 1 indicating that near surface air parcels with θ_e larger than 360 K are rare, with Figure 3 showing that radiatively diagnosed convective outflow above 15 km is weak, and with Figure 4 suggesting that the annual temperature variation at 16.85 km is under stratospheric control. There are however important regional variations in near surface θ_e. Analysis of high resolution radiosonde data from TOGA/COARE (not shown) indicates that $N(\theta_e)$ distributions at some sites within the Western Pacific Warm Pool are shifted by up to 5 K to the right of the tropical averages shown in Figure 1. In addition, the lidar measurements referred to earlier were taken during the season (Northern Hemisphere winter) and within the region (Western Pacific Warm Pool) where potential temperatures at the cold point tropopause are smallest. The near surface θ_e distribution and vertical temperature profile during Northern Hemisphere winter therefore conspire together to make the cold point tropopause over the Western Pacific Warm Pool more thermodynamically acces-

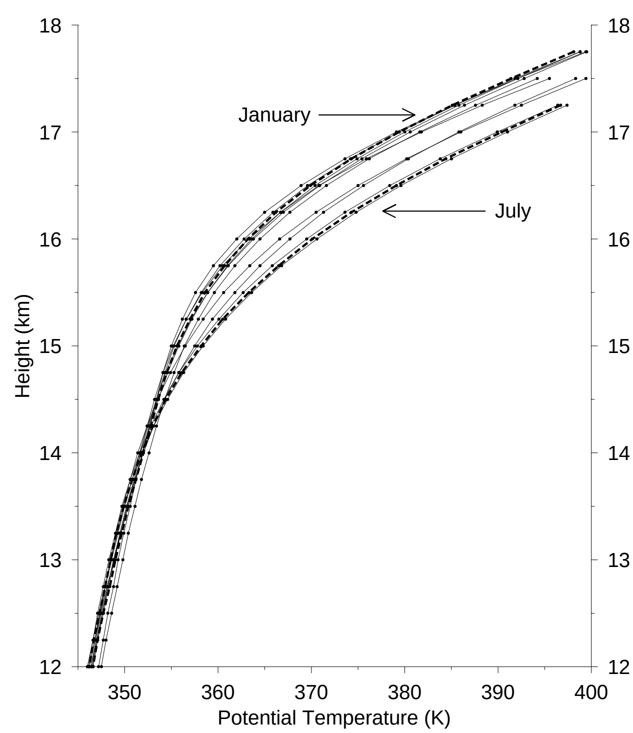

Figure 4. Monthly tropical (15 °S to 15 °N) climatologies of potential temperature versus height. The months of January and July are denoted by thick dashed lines.

sible to deep convection than is elsewhere generally the case [*Newell and Gould-Stewart*, 1981].

There are several indications that water vapor concentrations in the stratosphere have increased over the past several decades [*Oltmans and Hofmann*, 1995]. This development may have important climate implications [*Forster and Shine*, 1999]. One explanation is that tropopause cold point temperatures might have increased in response to surface warming [*Kirk-Davidoff et al.*, 1999]. Predicting the manner in which cold point temperatures will respond to global warming is made difficult by the fact that it is near a transition from radiative-convective (troposphere) to near radiative (stratosphere) equilibrium. Temperatures in the upper tropical troposphere may respond to surface warming in such a way as to maintain the scaling between upper tropospheric mass fluxes and the near surface θ_e distributions shown in Figure 2. If so, this constraint may give some insight into how temperatures in this important transition region can be expected to evolve.

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References

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

Dessler, A. E., A reexamination of the "stratospheric fountain" hypothesis, *Geophys. Res. Lett., 25*, 4165-4168, 1998.

Emanuel, K. A., Atmospheric Convection, Oxford University Press, 1994.

Folkins, I., M. Loewenstein, J. Podolske, S. Oltmans, and M. Proffitt, A barrier to vertical mixing at 14 km in the tropics: Evidence from ozonesondes and aircraft measurements, *J. Geophys. Res., 104,* 22095-22101, 1999.

Forster, P. M., and K. P. Shine, Stratospheric water vapor changes as a possible contributor to observed stratospheric cooling, *Geophys. Res. Lett., 26*, 3309-3312, 1999.

Fu, Q., and K. N. Liou, On the correlated k-distribution method for radiative transfer in nonhomogeneous atmospheres, *J. Atmos. Sci., 49,* 2139-2156, 1992.

Highwood, E. J., and B. J. Hoskins, The tropical tropopause, *Q. J. R. Meteorol. Soc., 124*, 1579-1604, 1998.

Kirk-Davidoff, D. B., E. J. Hinsta, J. G. Anderson, and D. W. Keith, The effect of climate change on ozone depletion through changes in stratospheric water vapor, *Nature*, 402, 399-401, 1999.

Newell, R.E., and S. Gould-Stewart, A stratospheric fountain?, *J. Atmos. Sci., 38*, 2789-2796, 1981.

Oltmans, S. J., and D. J. Hofmann, Increase in lower-stratospheric water vapour at a mid-latitude Northern Hemisphere site from 1981 to 1994, *Nature, 374,* 146 - 149, 1995.

Sassen, K., R. P. Benson, and J. D. Spinhirne, Tropical cirrus cloud properties derived from TOGA/COARE airborne polarization lidar, *Geophys. Res. Lett., 27*, 673-677, 2000.

Thuburn, J., and G. C. Craig, GCM tests of theories for the height of the tropopause, *J. Atmos. Sci., 54,*, 869-882, 1997.

Tompkins, A. M., and G. C. Craig, Radiative-convective equilibrium in a three dimensional cloud-ensemble model, *Q. J. R. Meteorol. Soc., 124*, 2073-2097, 1998.

Yulaeva, E., J. R. Holton, and J. M. Wallace, On the cause of the annual cycle in the tropical lower stratospheric temperature, *J. Atmos. Sci., 51*, 169-174, 1994.

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