Background Notes - Tropical Dynamics Section, Saturday August 9, 2003

This does not include the figures. They are referenced for the most part, but if you would like a copy please email me! (Ian.Folkins@dal.ca)

Global Chemistry and Climate of the Troposphere/Lower Stratosphere Summer School

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Section 1: Background Thermodynamics

We want to be able to address the following inter-related questions:

1) Is there a thermodynamic criteria that tells us whether or not an air parcel near the surface has sufficient temperature and/or moisture to allow it to participate in deep convection?

2) Is there a quantity that is conserved as an air parcel rises inside a convective cloud (assuming no mixing between air parcels in the cloud or with the outside environment).

3) What is the buoyancy acceleration experienced by an air parcel as it rises inside a cloud? To answer this question, we need to know the temperature and condensate loading of the air parcel.

4) Assuming no mixing or entrainment, at what height should an air parcel detrain? i.e. what is the level of neutral buoyancy of a surface air parcel? Can thermodynamics help explain "how high clouds go", i.e. where clouds deposit air? How is the LNB modified by entrainment, condensate loading, cloud radiative effects, freezing of condensate?

These questions are not really answerable at the current time. This is mainly due to a lack of knowledge of cloud microphysics inside convective systems. We are, however, able to give some answers to these questions in two different idealizations: reversible ascent and pseudoadiabatic ascent. The main success of these idealizations is that they appear to have some success in predicting the mean temperature structure of the tropical troposphere (Section 2).

Reversible ascent: This assumes adiabatic ascent (no exchange of heat or mass with the environment) of air parcels rising from the surface, and that all phase changes occur at equilibrium. In particular, all condensate is retained during the ascent. This is likely a good approximation in the lowest few km, but is probably a very poor approximation in the upper troposphere. Over the oceans, for example, convective updraft speeds are $\sim$ 5-10 m/sec, so large particles will not be lofted upward.

Pseudoadiabatic ascent: One assumes that condensate is removed as soon as it is formed. This means that the moist entropy of the air parcel is not conserved, but decreases due to loss of entropy of the condensate (due to the rapid decrease in saturated vapor pressure with temperature).

Some Differences:

(1) In reversible ascent, there is the potential for a large release of heat to the air parcel when it crosses the freezing level, and the lofted condensate freezes. This effect is much smaller in the irreversible process since only a fraction of the original water vapor will remain in the air parcel by this point.

(2) Buoyancy effects: If the condensate is lofted upward this will decrease the buoyancy of the air parcel. You decrease what is called the "Density Temperature" T_ρ of the air parcel.

(3) Retaining the condensate makes the heat capacity of the air parcel bigger, so slows down temperature decrease during ascent, like a "thermal memory" of surface warmth.

To some extent, these effects offset one another, so the differences are not as big as you might think. See: Williams and Renno, MWR 121, p. 21, 1993. Xu and Emanuel, MWR, 117, 1471 - 1479, 1989.

CAPE: This is the work done on an air parcel (J/kg) as it rises from the Level of Free Convection (LFC) to the Level of Neutral Buoyancy (LNB).

$$CAPE = \int_{LFC}^{LNB} (T_\rho - T_b) R_d dlnP, \tag{1}$$

where, T_b is the temperature of the background atmosphere, and T_ρ the density temperature of the ascending air parcel. CAPE is a function both of the temperature/moisture of the air parcel at the surface and the background temperature of the atmosphere, and also of course depends on whether you assume reversible or pseudoadiabatic ascent. In a colder atmosphere, upward buoyancy forces on an air parcel are larger and CAPE will be bigger. Near the equator, however, free tropospheric temperatures are quite homogeneous, and so there is a very strong correspondence between CAPE and θ_e, with CAPE becoming positive for θ_e larger than ~ 344 K. The main usefulness of the CAPE concept is that it tells us which air parcels near the surface are thermodynamically able to participate in deep convection, and also in that it gives us an upper limit on updraft speeds one would expect to encounter. Positive CAPE air parcels in the tropics generally have a range in CAPE of 0 - 2000 J/kg. Since the vertical velocity at the LNB is $v = \sqrt{2CAPE}$, this would lead to updraft speeds of 0 - 63 m/sec. However, observed updraft speeds over the ocean are only 5 - 10 m/s. Differences are usually ascribed to condensate loading or entrainment, but the relative roles played by the two processes have not been fully characterized.

θ_e **equivalent potential temperature**: a quantity that is conserved during moist adiabatic ascent and reduces to θ when the RH is zero. It is a function of pressure, temperature, and relative humidity.

θ_{ep}**: Pseudo-equivalent potential temperature:** the potential temperature that is realized by lifting an air parcel upward along a moist pseudoadiabatic trajectory, extracting all the water vapor, and returning adiabatically to the surface (i.e. $\theta_{ep} - \theta$ is how much "extra temperature" the air parcel would have, at STP conditions, if all the moisture were condensed via the pseudoadiabatic pathway.)

θ^*_{ep} **Saturated pseudo-equivalent potential temperature**. The θ_{ep} an air parcel would have if it were saturated. It is just a function of temperature and pressure. θ^*_{ep} - θ_{ep} is in a sense a measure of relative humidity. Air parcels rising inside updrafts are saturated, so they will have a LNB where their θ_{ep} is equal to the θ^*_{ep} of the background atmosphere (ignoring virtual temperature effects and the effects of condensate loading).

Detrainment: This refers to an air parcel exiting a cloud (usually defined as the point at which there is no condensate).

Entrainment: This refers to an air parcel being mixed into a cloud from the ambient atmosphere. Also used in the context of air parcels entrained into a convective downdraft (plumes of sinking air caused by evaporation from ice/water).

Figure 1: (Taken from the CEPEX design document, *Ramanathan*) This shows the overturning circulation in the tropics. Air rises up to the tropopause (up to 17 km) inside relatively narrow updrafts, which then spread out into huge cirrus anvils covering 100's of km. Air then radiatively subsides ($\sim$ 1 km/day) in large clear sky areas, often over the colder SST regions, which are covered by low ($\sim$ 2 km) stratocumulus. The air is then advected to warmer SST regions, picking up heat and moisture, until it attains positive CAPE, and can then repeat the

cycle. In addition to the deep convective and stratocumulus clouds, there is an intermediate category of cumulus congestus with cloud tops up to the melting level ($\sim$ 5 km). The upward mass flux from the cumulus congestus, or mid-level, clouds appears to be balanced, to first order, by downward mass fluxes of evaporative downdrafts (not shown). These mass fluxes are about one-third of the total, or half as big as the deep convective mass fluxes (balanced by radiative descent). They give rise to a local, mid-tropospheric circulation of air within the warm SST actively convecting regions.

Figure 2: (from *Mapes*, Q. J. R. Meteorol. (2001), 127, pp. 2353 - 2366) One starts off with an air parcel at the surface with $\theta_e = 351$ K. The thick solid line is the temperature difference between an air parcel lifted upward along a moist reversible adiabat and a background tropical temperature. The thick dashed line is the difference between the density temperature (T_ρ) of this adiabat and the background temperature ($T_\rho = T\frac{1+r/\epsilon}{1+r_T}$, where r is the water vapor mass mixing ratio, r_T the total water mass mixing ratio, and $\epsilon = 0.622$). The density temperature gives the amount by which the buoyancy acceleration of an air parcel is reduced due to the presence of the condensate (i.e. $B \propto (T_\rho - T_b)$). Note the large increase in buoyancy as the parcels rise up through the melting level and the condensate freezes. The thin solid line is the $(T - T_b)$ of a pseudoadiabat. The thin dashed line is the $(T_\rho - T_b)$ of a pseudoadiabat. In this case, there is no condensate, so there is only a slight increase in buoyancy due to virtual temperature effects (i.e. the fact that water vapor has a smaller molecular mass than air). An important point of this figure is that, even though neither the reversible or pseudoadiabatic ascent idealizations can be expected to be entirely realistic, their predictions for density temperatures are quite similar.

Figure 3: θ_e as a function of temperature and relative humidity.

Figure 4: (From Bohren and Albrecht, Atmospheric Thermodynamics) The thick solid line is the temperature of the background atmosphere; the line with arrows the temperature of an air parcel undergoing moist adiabatic ascent. LCL (Lifting Condensation Level): where the relative humidity of the rising air parcel reaches 100 %; LFC (Level of Free Convection): where the temperature of the air parcel becomes warmer than the environment. The light line is the dew point temperature of the background atmosphere.

Figure 5: CAPE as a function of θ_e at a typical tropical location. CAPE is undefined if

not positive.

Figure 6: Two moist pseudoadiabats and a mean tropical temperature profile.

Figure 7: θ, θ_{ep}, θ^*_{ep} at Koror. Note that they all become essentially identical above 12 km. This is the height at which, owing to the very cold temperatures, the contribution of water vapor to the moist entropy, or moist static energy, of the air parcel becomes negligible.

Figure 8: θ_e, RH, and temperature in the Boundary Layer at Koror. Heat and moisture and are mixed upward from the surface to the top of the boundary layer on a timescale of $\sim$ 1 day, or less. The top of the BL is characterized by sharp gradients in RH, a decrease in θ_e, and in increase in stability.

Standard References: *Atmospheric Convection* by Kerry Emanuel, and *Atmospheric Thermodynamics* by Bohren and Albrecht.

Section 2: Mean Vertical Temperature Structure

Figure 1: This figure shows that mean temperature profiles are quite similar at tropical locations. This is apparently due to the low value of f, which makes pressure gradient forces (as would arise from temperature anomalies) difficult to sustain, so that heating and cooling simply give rise to ascent and descent (to first order). This is an important simplification and seems to have profound implications for how rainfall is distributed in the tropics (more below). The cold point tropopause occurs at 17 km, and in actively convecting regions, is the upper limit of moist convection. However, only a tiny fraction (probably less than 1%) reaches this height, with most detraining below 14 km. The transition to clear sky radiative heating occurs near 15 km, so the amount of convection that detrains above this level may be roughly equal to the amount needed to sustain the stratospheric Brewer-Dobson circulation.

Figure 2: The tropical troposphere is often described as having a moist adiabatic temperature structure. That is, that the mean temperature profile is approximately equal to the temperature profile one would generate by starting with an air parcel at the surface, and lifting it upward along a moist adiabatic, or pseudoadiabatic, trajectory. The dashed line in this figure is the lapse rate generated by a 345 K moist pseudoadiabat. Observed lapse rates (here dT/dz) in strongly convecting regions most closely approximate the dashed line between about 5 and 12 km. Subtropical locations like LIH and HIL (in Hawaii) have strong boundary layer inversions, and lapse rates that are not close to moist adiabatic at any height. The physical mechanisms

responsible for the deviations from a moist adiabat below the 5 km melting level have not been fully explained, but probably have to do with the greater prevalence of convective downdrafts in these regions. (Unsaturated downdrafts would tend to drive the atmosphere to a lapse rate that is not moist adiabatic.)

Figure 3: Annual mean tropical temperatures at 14 different locations were averaged together to make a tropical mean. These plots show the deviation of each station from the tropical mean. Stations characterized by active convection are warmer than the mean below 14 km, but cooler than the mean above 14 km. Differences are mostly less than 2 K below 14 km.

Figure 4: Each curve is the JJA - DJF temperature difference at each location. The amplitude of the seasonal cycle increases rapidly above 15 km. Temperatures in the tropical lower stratosphere are lower during Northern Hemisphere winter due to stronger upwelling from the Brewer Dobson circulation.

Section 3: Nonlinear Dependence of Rainfall on SST

Figure 1: Shows the climatological SST patterns in the tropics, together with the climatological rainfall pattern.

Figure 2: Each dot is the rainfall rate for a month in 1991 plotted against the SST for that month over an 2.5 by 2.5 degree area of the ocean. There is obviously a lot of scatter. But there is a annual mean relationship that shows a rapid increase in rainfall near 27 °C, and then a decrease at very high SST's.

Figure 3: Similar to Figure 2, but shows 10S-10N averages.

Figure 4: This figure is a PDF showing the amount of area at each SST in the tropics (30 S - 30 N). It is also a highly skewed distribution, with a rapid cut off of SST's greater than about 30 °C.

Figure 5: This figure is taken from *Folkins*, J. Clim., 1 June 2003, where I argued that the nonlinear dependence of rainfall on SST in the tropics can be related to the nonlinear dependence of something called "convective mass" in the tropics. This figure tries to demonstrate how convective mass is calculated. Each dot corresponds to a θ_{ep} measurement below 800 mb ($\sim$ 2 km) above Koror, a location within the Western Pacific Warm Pool. The vertical dashed line is the approximate location of the convective threshold θ_e, i.e. the θ_e at which the CAPE of near surface air parcels becomes positive. The points to the right of the dashed line would

contribute to the convective mass, while those to the left would not. The solid line is the mean θ_{ep} for that month.

Figure 6: The monthly mean convective mass at various tropical locations was calculated, and is plotted here against the SST at that station. If the convective mass is 100 mb, it means roughly half the mass at that station has positive CAPE, in a time average, below 800 mb (the contribution above 800 mb is small). There is a significant increase in convective mass at an SST of 27 C.

Figure 7: This is similar to Figure 6 but shows convective water rather than convective mass. Convective water is the depth of the water column you would get by condensing out all the water vapor in the convective mass. One can think of it as the maximum amount of rain that could be generated locally from the moisture present in the boundary layer. These are monthly means. Instantaneous values could be higher. The dashed curve shows the mean tropical rainfall versus SST relationship.

Figure 8: This figure tries to give a heuristic explanation for why there appears to be an abrupt separation of the tropics into weakly and actively convecting regimes, along the lines suggested by the above. The upper panel shows the result of a thought experiment where one imagined that there were rigid walls between SST regions in the tropics, so that within each SST, convection was balanced by subsidence (see *Emanuel*, QJRMS, 120, pp 1111, 1994). In this case, the higher SST regions would give rise to a warmer atmosphere above them. Within each box, the production of positive CAPE air via heat/moisture fluxes from the ocean would be balanced by the rate at which atmospheric cooling supplied low θ_e air to the boundary layer. Rain would probably only increase moderately with SST (reflecting the slow increase in clear sky radiative cooling with SST), and the convective mass would be similar everywhere. If you remove the walls, an atmospheric circulation would develop which tried to homogenize atmospheric temperatures (and the convective threshold θ_e). The increase in temperature above the low SST regions would suppress the convective mass in these regions, and reduce rainfall. The water vapor evaporating off the colder SST regions would instead be advected to the warmer SST region, by a circulation balancing the mass transport at the upper levels. The sharpness of the transition between the weakly and actively convecting regimes is therefore perhaps mainly a reflection of the efficiency at which the tropical circulation redistributes heat

above the boundary layer. This highly efficient long range transport also makes the prediction of tropical rainfall patterns a highly nonlocal problem. For example, during ENSO events, there is often a suppression of rainfall over the Atlantic. This could be attributed to a transient increase in the convective threshold θ_e over these regions, due to the increase in mean SST's in the Pacific.

Thermodynamics versus Dynamical Explanations for Rainfall: Convection can only occur given the presence of air in the Boundary Layer with positive CAPE. But the presence of positive CAPE air near the surface is extremely common in the tropics, so thermodynamics only provides a precondition, and probably can't help predict the evolution of convective systems on short timescales. However, thermodynamics does appear useful in explaining large scale patterns in the distribution of rainfall. The role of dynamics in organizing tropical rainfall is explored in the next section. Obviously, one should always pay attention to both aspects of the problem, as there is unlikely to be any strict separation at any given time/spatial scale.

Section 4: Convectively Coupled Tropical Waves

Wheeler, M. C., *Equatorial Waves*, in "Encyclopedia of Atmospheric Sciences" (Academic Press, edited by Holton et al.)

Wheeler and Kiladis, JAS 56, pp 374 - 399, 1999.

Mapes, JAS 52, pp. 1807 - 1828, 1995.

Figure 1: (Figure 8 from *Wheeler and Kiladis*) This plot shows OLR in the tropics. There is nearly continuous convection over Africa (20 W) and Brazil (60 W). There is much less convection over the ocean between Brazil and Africa and over the colder SST region in the eastern Pacific. In the Indian and Western Pacific there is both eastward and westward propagating convection. Eastward is down and to the right, westward is is down and to the left. Propagation that follows an almost horizontal line is "fast".

Figure 2: (Figure 9 from W and K) This identifies the types of waves in Figure 1. The large eastward moving "blobs" of convection are the Madden Julian Oscillation (MJO). Also eastward are the Kelvin waves, while Equatorial Rossby (ER) and Mixed Rossby Gravity (MRG) waves are westward propagating.

Figure 3: (Fig 3 of W and K). This shows the power from 18 years of data at various wave numbers and frequency. The existence of convectively coupled waves is revealed by maxima

along particular dispersion curves. These curves arise from the equatorialy trapped solutions to the shallow water equations on the β plane, linearized about a basic state of no motion (see the Wheeler encyclopedia article for more details. This sounds very complicated, but keep in mind that any fluid medium will have normal modes of motion, depending on its geometry, stratification, basic state, etc. It is perhaps surprising that the modes that can be found analytically under the above approximations end up being of such relevance to tropical meteorology.) The curves for the various types of waves have been labeled with equivalent depths values h of 12, 25, and 50 m. The equivalent depth "appears in the shallow water equations, and is simply the depth of the shallow layer of liquid required to give the correct horizontal and time-varying structure of the mode." (Wheeler). The equivalent depth that appears to best describe the observed dispersion curves is $h = 25$ m. Why is this?

There are mathematical methods of determining h given the depth of the heat source that initiates the wave, and the tropical mean temperature profile in which the heat source is embedded (see W and K for more details). A heat source extending the full length of the troposphere from 0 to 14 km (as one might expect from deep convection) would be considered to have a wavelength $L_z = 28$ km. For a mean $dT/dz = -7k/km$ and a mean temperature of 250 K, you would generate waves with $h = 200$ m (W and K). h scales approximately with the square of L_z. The equivalent depth h is related to the internal gravity wave speed as $c = \sqrt{gh}$. From Figure 3, the equivalent depth which characterizes the dispersion curve with the most power has $h \sim 25$ m. The discrepancy between these two depths - i.e. the depth you would expect to be associated with a deep convective heat source giving rise to upward vertical motion extending from the surface to the tropopause (200 m), and the real equivalent depth of convectively coupled waves (25 m) - has been a "mystery" ever since the dispersion curves of equatorialy trapped waves were predicted in the late 60's, and has inspired many different solutions (e.g. moist convective damping). That is, why are convectively coupled waves slower than you would expect? Of course, the fast waves are observed in the tropics in the dynamical fields - it's just that they do not initiate convection as they propagate.

I will discuss a proposal first given in Mapes (1995), which seems to have the most potential, and explain the phenomenon most naturally. As noted earlier, the spectrum of waves with various equivalent depths (or phase speeds) that is excited by deep convection can be determined from the background temperature profile, and the heat source as a function of height. The strategy used by Brian Mapes was to use cloud radar data to effectively determine the vertical variation of deep convective heating. Mapes (1995) wrote the thermodynamic equation as,

$$C_p \frac{dT_h}{dt} - \omega\sigma = Q$$

where C_p is the heat capacity at constant pressure, T is temperature, ω is the vertical velocity, σ is a static stability parameter ($\sigma = (\Gamma_d - \Gamma)/\rho g$), Q the heating rate (arising from latent heat release, turbulence, radiation inside a cloud) and d_h/dt refers to local change plus horizontal advection. In the tropics, to first order, heat sources give rise to vertical motion rather than changes in the local temperature, so that,

$$\omega \sim -Q/\sigma.$$

It is difficult to determine the vertical motion field inside a cloud from observations. If, however, you can measure the horizontal velocity field surrounding a convective system, you can determine the export of mass from the convective system, which gives you the divergence δ. This is related to the average ω inside the cloud via

$$\delta = (\bigtriangledown \cdot V_h) = \frac{\partial \omega}{\partial p}.$$

I am simplifying the argument - see his paper for more details. Once a mean δ is compiled from many convective systems (this is tricky because δ evolves over the lifetime of a system), you can decompose this is into its vertical mode spectrum. δ just represents the overall mass redistribution that has to be accomplished by some combination of gravity waves. Due to the particular vertical shape of δ and the temperature stratification of the atmosphere, waves of particular phase speeds (or vertical wavelength) play a much bigger role than some others.

Figure 4 (Figure 11 of *Mapes*, 1995): This shows the mean divergence profile δ.

Figure 5 (Figure 14 of *Mapes*, 1995): This shows the vertical mode spectrum of the mean δ profile. Note the maximum at $c_n = 50$ m/sec. This corresponds to $h \sim 250$ m and arises from the deep heating, or convective, part of the mean δ profile. Waves with these phase speeds are frequently observed in the tropics but are not coupled to convection. There is another peak at $c_n = 24$ m/sec, which is associated with mid-tropospheric convergence near the melting level driven by evaporative cooling.

The lower part of this figure shows the temperature perturbations associated with this wave spectrum. The deep heating response travels outward very fast - i.e. 1000 km after six hours, so $c \sim 46$ m/sec. This heating throughout the whole troposphere would be expected to be associated with a suppression of convection, since a warm atmosphere reduces the CAPE of near surface air parcels, and would reduce the convective mass. This appears to explain why the fast response is not convectively coupled. The slower mode with $c \sim 24$ m/sec gives rise to a cooler lower troposphere and a heated upper troposphere. The cooler lower troposphere would increase the convective mass in the boundary layer below it, and make convection more thermodynamically favorable. The cooling would also be associated with large scale uplift, which could initiate convection by low level convergence. This appears to be the explanation for why the convectively coupled waves travel at the slower phase speed.

Weather forecast models and GCM's continue to have difficulty in reproducing the propagation of convectively coupled tropical waves. In the case of forecast models, these difficulties may be important in limiting the accuracy of medium range forecasts. This is partly an initialization problem. However, from the point of view of modeling, the above explanation for the coupling of equatorial waves and convection, would suggest that models should try to make sure they have the correct δ from deep convective systems (of obvious importance would that component driven by evaporative cooling), that this divergence generates an appropriate spectrum of waves, and that the temperature perturbations initiated by these waves help trigger or inhibit convection in the boundary layer beneath them in a realistic manner. Convective propagation will also be affected by the low level circulations associated with these waves.

It is interesting to speculate how the mean δ of convective systems might change in warmer climate, and the effects this might have on the propagation of convectively coupled waves. In particular, the phase speed of the slower mode may be influenced by the height of the melting level, which is known to be increasing.

MJO: It is now believed that the MJO is a separate beast, i.e. that it is not an equatorialy trapped wave that is simply modified by convection. Part of this reasoning comes from the fact that it lies off the Kelvin wave dispersion curve (Figure 3). Some GCM's have been able to generate MJO-like oscillations. However, a fundamental explanation of their origin and propagation speed continues to be an outstanding problem in tropical meteorology.