Radiative and Convective Driving of Tropical High Clouds

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Abstract

Using satellite cloud data from Aqua MODIS and collocated precipitation rates from Advanced Microwave Scanning Radiometer (AMSR), it is shown that rainrate is closely related to the amount of very thick high cloud, which is a better proxy for precipitation than OLR. It is also shown that thin high cloud, which has a positive net radiative effect on the top-of-atmosphere (TOA) energy balance, is nearly twice as abundant in the West Pacific compared to the East Pacific. For a given rainrate, anvil cloud is also more abundant in the West Pacific. The ensemble of all high clouds in the East Pacific induces considerably more TOA radiative cooling compared to the West Pacific, primarily because of more high, thin cloud in the West Pacific. High clouds are also systematically colder in the West Pacific by about 5 K.

We examine whether the anvil cloud temperature is better predicted by low-level equivalent potential temperature ($\Theta_E$), or by the peak in upper-level convergence associated with radiative cooling in clear skies. The temperature in the upper troposphere where $\Theta_E$ is the same as that at the lifting condensation level (LCL) seems to influence the temperatures of the coldest, thickest clouds, but has no simple relation to anvil cloud. We show instead that a linear relationship exists between the median anvil cloud top temperature and the temperature at the peak in clear-sky convergence. The radiatively-driven clear-sky convergence profiles can thus explain why anvil cloud is warmer in the East Pacific compared to the West Pacific.
1. Introduction

Upper-tropospheric clouds associated with tropical convection have large shortwave (SW) and longwave (LW) cloud radiative forcing (CRF). Clouds reflect SW radiation, which cools the planet, and the size of the effect is determined primarily by cloud optical depth. Clouds trap LW radiation, which tends to warm the climate, and primarily depends on cloud top temperature, except for relatively thin clouds (visible optical depth $\tau \leq 4$), in which the LW effect increases with optical depth as well. The ensemble of convective clouds is such that the SW and LW effects come fairly close to canceling (Harrison et al. 1990). Using International Satellite Cloud Climatology Project (ISCCP) data, Hartmann et al. (2001) show that in an area of the West Pacific, the net CRF is on the order of about -10 Wm$^{-2}$, and a considerably larger negative net CRF in a portion of the East Pacific ITCZ region of about -45 Wm$^{-2}$ is observed. It is suggested that this is because clouds are warmer and optically thicker in the East Pacific. Berg et al. (2002) and Schumacher and Houze (2003) have also studied structural differences, and infer the existence of shallower rainfall systems with more stratiform precipitation in the East Pacific.

We construct temperature-optical depth (T-τ) histograms to better understand differences in high cloud properties in the West Pacific (WP) (5°-15°N, 120°E-160°E), Central Pacific (CP) (5°-15°N, 160°E-160°W), and East Pacific (EP) (5°-15°N, 150°W-100°W). These histograms are then used in conjunction with a radiation model in order to quantify radiative effects of convection across the ITCZ, in a similar spirit as Hartmann et al. (2001), but with much higher vertical resolution afforded by MODIS data. We also are interested in the driving mechanisms behind the structure of these histograms, and thus we composite high cloud fraction with rainrate. Composites by rainrate normalize convective intensity, so that we can then examine
more subtle effects on convection, such as those from SST, SST gradients, and ITCZ structure across the North Pacific.

We also wish to address the claim that cloud systems are shallower in the EP, but in doing so we ask the question how high tropical clouds get and what controls their detrainment level. Reed and Recker (1971) analyzed satellite data composites that revealed a peak in stratiform anvil and thin cloud around 175 hPa, which corresponds to approximately 13 km. They also made upper-level divergence calculations, which suggested a peak at around the same level. Other studies have also suggested convective detrainment well below the tropopause. Fokins et al. (1999, 2000) suggest that convective detrainment occurs below 14 km, based on O₃ concentrations that begin increasing towards stratospheric concentrations at this level.

So, what determines the top of the tropical convective layer, and why is this top apparently well below the tropopause? The 'Hot Tower' concept of Riehl and Malkus (1958) and Riehl and Simpson (1979) assume that convection occurs such that saturated ascending parcels are undiluted, and become negatively buoyant around 150 hPa, making this a cap for convection, which again is below the tropopause. Fokins et al. (1999 & 2000) also state that undiluted parcels become negatively buoyant once they reach the level in the upper atmosphere in which environmental $\Theta_E$ is the same as $\Theta_L$ at the LCL. We refer to these ideas as the 'PUSH' concept, as high clouds are convectively driven by energy present in low-level air. A contrasting perspective, which we refer to as the 'PULL' concept, is that clear-sky radiatively-driven upper-level convergence determines the level at which detrained cloud is maximum. Clear-sky radiative cooling drops off quickly in the upper troposphere where water vapor levels rapidly decrease, since saturation vapor pressure is dependent only on absolute temperature, per the
Clausius-Clapeyron relationship. Hartmann and Larson (2002) have argued that the PULL mechanism implies an essentially fixed anvil cloud temperature.

In order to evaluate the PULL method, we use a radiative transfer model to calculate clear-sky cooling rates and the corresponding upper-level convergence. We then examine to what extent anvil outflow occurs at this level, as evidenced by cloud temperatures measured by MODIS. We will quantify the relationship between clear-sky convergence and cloud temperature. The goal is to characterize the mechanisms that determine the altitude of convective clouds in the ITCZ.

2. Data for Cloud T-τ histograms

We use data from Aqua MODIS, which is a sun-synchronous satellite with local daytime overpass time at the equator of 1:30 p.m. The MODIS cloud products include cloud mask (which assesses the probability that there is a cloud in a given pixel), cloud top properties (i.e. temperature, pressure), cloud optical depth, and cloud phase. The horizontal spatial resolution of five km of the Joint Level-2 MODIS dataset is higher than other remote-sensing instruments of similar spatial coverage (Platnick et al. 2003). The nature of a sun-synchronous satellite, however, does preclude the observation of the diurnal cycle.

Cloud optical depth retrievals are based on solar reflectance. Clouds are assumed to be homogeneous, and ice and liquid cloud look-up tables are used (King et al. 1997). A non-absorbing wavelength band (0.86 µm) is used to reduce the effect of particle size (King et al. 1997).

To estimate cloud top temperature, the Joint Level-2 dataset is used, which has the same horizontal resolution for cloud top temperature and cloud fraction (5 km) as the full Level-2 data. Though the Joint dataset has lower resolution for τ (5 km versus 1 km) than the standard Level-2
swath product, its smaller data volume is an advantage. For clouds above 700 hPa, a CO$_2$-slicing method is used (Platnick et al. 2003). This method utilizes the partial absorption in each of the different MODIS infrared bands that are located in the CO$_2$ 15-µm absorption region, where each MODIS band senses a different atmospheric layer. Upward infrared emission is measured in each of the different bands, and cloud pressure is then determined. With use of the National Center of Environmental Prediction (NCEP) reanalysis tables, cloud top temperature is then estimated. In the lower troposphere (below 700 hPa), the CO$_2$-slicing method is less effective, and is replaced with band temperatures within the 11-µm atmospheric window. One major advantage of CO$_2$-slicing is that it is relatively insensitive to emissivity, so that the temperature of thin clouds can be sensed more accurately compared to estimates based on brightness temperatures alone (Platnick et al. 2003).

Because our interest is in tropical oceanic convection, only areas over the open ocean are considered. Also, because we are interested in both thermal and optical properties, only daytime data are used. Data influenced by sunglint are not used.

Instantaneous satellite data are aggregated into three-day averages and 1° latitude by 1° longitude regions. This averages together individual convective events occurring on short time (less than one day) and small spatial scales (less than about 100 km). These temporal and spatial scales are small enough to capture variability in convection.

We use weekly sea-surface temperature (SST) data from the Climate Diagnostics Center, which contain National Oceanic and Atmospheric Administration (NOAA) Optimum Interpolation (OI) SST version two data. These data are global, have 1° x 1° resolution, and are derived from a combination of in-situ and satellite measurements (Reynolds et al. 2002). These data are interpolated onto the same temporal scale as the MODIS data.
Daily rainrate data on a grid of 0.25° by 0.25° comes from the Advanced Microwave Scanning Radiometer (AMSR), which is also aboard the Aqua satellite. The obvious advantage of this is the collocation (in space and time) with MODIS. Its microwave frequencies can see larger raindrops through smaller cloud particles, and it becomes saturated for precipitation rates of 25 mm/hr. We use the version five rainrates, which have an improved rain algorithm in which freezing levels have been revised, lowering calculated rainrates in the tropics compared to older versions (Updates to AMSR-E V05 Algorithm 2006). Like the SST data and MODIS data, rainrate data are aggregated into three-day averages for compositing purposes with MODIS cloud data.

To further ensure the statistical robustness of the MODIS data, we use only aggregated three-day 1° latitude by 1° longitude data that contain at least 200 good 5km boxes. During a three-day period, a maximum of 1200 good boxes are possible, since 1° of latitude/longitude near the equator is about 100km, and the horizontal resolution of Joint Level-2 MODIS is 5km. The choice of 200 granules is somewhat arbitrary, but does filter out averages that contain very little good data, whether from sunglint, proximity to land, or other data quality issues.

To understand both qualitatively and quantitatively the structure of clouds across the ITCZ, cloud fraction histograms, categorized by temperature (ordinate) and optical depth (abscissa) are constructed, which we will refer to as T-τ histograms. The three regions, namely the WP, CP, and EP have different SST distributions, with the WP representing the large warm pool region, the CP a transition zone of sorts, and the EP a narrow ITCZ, with strong SST gradients on its edges. A composite SST map from September 2003 through August 2005 is given in Fig 1. It should be noted that the median SST in the WP during this period is 28.9°C, 28.5°C in the CP, and 27.9°C in the EP. For the T-τ histograms, 26 temperature bins of five
degrees and 10 optical depth bins are used, the latter of which follows a nearly logarithmic scale with the following bins: (0-0.125, 0.125-0.25, 0.25-0.5, 0.5-1.0, 1.0-2.0, 2.0-4.0, 4.0-8.0, 8.0-16.0, 16.0-32.0, 32.0-64.0). We have a total of 260 possible cloud types, which provides much greater resolution than the 42 ISCCP pressure/optical-depth bins. However, it is primarily in the vertical that MODIS T-τ histograms are superior to ISCCP. For example, in the upper troposphere, the ISCCP pressure bins are 440 hPa to 310 hPa, 310 hPa to 180 hPa, and 180 hPa to 30 hPa. These correspond to approximate temperature ranges (based on GPS data) of 260 K to 240 K, 240 K to 215 K, and 215 K to 193 K (at 100 hPa), which are coarse compared to the 5 K bins used here. These histograms are discussed in a later section.

3. Radiative Transfer Model/T-τ Histogram Methodology

We employ the same radiation model as used by Hartmann et al. (2001), namely the Fu and Liou (1993) delta-four-stream, k-distribution scheme. The purpose of a radiative transfer model is twofold; we calculate radiative energy budgets for each of our cloud categories, and we also use clear-sky heating rates (to be discussed later) for calculation of clear-sky convergence profiles. We focus on the former in this section.

For each of the aforementioned cloud types, cloud radiative forcing is calculated, assuming 100% cloud cover in each bin. These are computed every hour for the centers of the WP, CP, and EP, with zenith angles corresponding to day 90 of the year. Ice water content (IWC) is calculated by the relationship

$$IWC = \frac{2}{3} \frac{r_{ice} \tau_{ice}}{\Delta z}$$

(1)

where $r_{ice}$ is the effective radius of ice, $\tau_{ice}$ the optical depth, and $\Delta z$ the geometric thickness of the cloud. An analogous relation is used for liquid water content (LWC); cloud becomes all ice
at temperatures below 263 K. No mixed clouds are included, but this is not seen as a major problem, since the emphasis of this study is that of tropical high cloud. Also, as is done in Hartmann et al. (2001), $r_{\text{ice}}$ is assumed to be 30µm, and $r_{\text{liquid}}$ 10µm. Geometric cloud thickness $\Delta z$ is based on climatology from Liou (1992), which contains nine thickness categories, depending on cloud height and optical depth. Concentrations of CO$_2$, CH$_4$, and N$_2$O are assumed to be 330 ppmv, 1.6 ppmv, and 0.28 ppmv, respectively, and the surface albedo is 0.05 (a reasonable value for the ocean).

In Fig. 2, contours of shortwave, longwave, and net radiation are shown in T-$\tau$ coordinates. Optical depth matters more for SW cloud forcing, and cloud top temperature has only a marginal effect, so that clouds have a much stronger SW cooling effect as their optical thickness increases. On the other hand, cloud top temperature and cloud optical depth are both important for LW cloud forcing for optically thin clouds ($\tau<4$ or so). For such clouds, as optical thickness increases, so does the LW warming effect. For all clouds, the LW effect increases with decreasing cloud top temperature, and for clouds with an optical thickness greater than four, only cloud top temperature matters. The net CRF is positive for thin cloud colder than 260 K, and reaches a peak (of +94.3 W m$^{-2}$) for very cold cloud at around $\tau=1$, and then decreases with increasing $\tau$. It becomes negative for larger optical depths, but becomes negative at lower optical depths for warmer clouds. We choose our high cloud to include clouds colder than 245 K, as this represents a temperature at which high cloud amount is well separated from other cloud modes (a relative minimum in cloud amount appears at approximately this temperature). Low, thick clouds have a negative forcing with a value of about -216 Wm$^{-2}$.
4. Cloud Fraction Histograms and Radiative Flux Histograms

a. Histogram cloud types

Cloud fraction histograms are presented in Fig. 3 for the period September 2003 through August 2005, and suggest three modes of clouds in the WP, CP, and EP. One mode is that of low, marine boundary layer cloud, with top temperatures generally warmer than 280 K. As described by Sarachik (1978), these have bases approximately at 600m that extend vertically to about 2 km. Yanai et al. (1973) note that the vast majority of the moisture is contained within this shallow cloud layer, and water vapor and liquid water are transported upward via deep convection. Another dominant mode is that of high cloud, which shows optical depths with a wide range of values. The observation of high cloud of a wide range of optical thickness, ranging from thick convective cores to anvil clouds to thin cirrus, is consistent with observations from ISCCP in Hartmann et al. (2001). Subtle differences exist in this cloud type, however, namely that high cloud peaks slightly higher up in the troposphere (lower cloud top temperature) in the WP and CP compared to the EP. Also, the median optical depth of all high cloud in the EP is slightly higher.

Mid-level clouds represent a third, though arguably less important mode, that are particularly prevalent in the EP. This cloud peaks around 260 K in the EP, and at slightly lower temperatures in the WP and CP. Less attention has been drawn to this population of clouds, and some studies, particularly more conceptual ones, have primarily underscored the two aforementioned modes. Early observational studies, however, such as that by Malkus and Riehl (1964), documented the presence of cumulus congestus clouds at or slightly above the 0°C level. Later studies, such as that by Johnson et al. (1999), highlight this middle population of clouds as an important one. Even a two-dimensional cloud modeling study by Liu and Moncrieff (1998)
reveals three cloud modes, one of which is located between about six to seven km, coincident with the melting layer.

To explain the reasons for these three modes, Johnson et al. (1999) suggest that the three layers; one near 2 km, another near 5 km (near 273 K), and the third in the vicinity of 15-16 km (about 1-2 km below the tropopause) represent three prominent stable layers of the tropical atmosphere (see Fig. 4 for GPS temperatures, corresponding heights, and differences between WP and EP). The enhanced stability in these areas stunts vertical cloud growth and promotes divergence and detrainment. In MODIS, we see the middle congestus clouds peaking considerably colder than 273 K, at 255 K to 260 K. Johnson et al. (1999) suggests that overshooting above the 0°C level occurs because supercooled droplets are in abundance, especially in the ITCZ, and droplets do not become glaciated until 10-15 K colder than 273 K. This is quite consistent with the MODIS identification of a middle cloud population colder than the 273 K level. It is less clear, however, why the population of congestus is more prominent in the EP. It may have to do with the fact that SST gradients are stronger in the EP, driving stronger surface convergence, and a profile of vertical velocity such that vertical velocity peaks much lower in the troposphere compared to regions such as the WP, where the SST gradients are relatively weak (Back and Bretherton 2006). Even when we examine the conditional probability of seeing congestus cloud given no high cloud (plot not shown), the EP still has more congestus cloud for a given rainrate.

b. Cloud forcing by type

Earlier, we presented T-τ histograms of SW, LW and total cloud radiative forcing, assuming 100% cloud cover in each T-τ bin. We now use the T-τ cloud fraction histograms constructed from MODIS data to calculate average-sky cloud forcing (the term used by
Hartmann et al. 2001), to provide information about the actual radiative impact of clouds for the WP, CP, and EP. Multiplying the corresponding overcast cloud forcing histograms (Fig. 2) with actual cloud fraction in each T-τ box (Fig. 3) gives the average cloud radiative effect.

We define high thin, anvil, and high thick cloud based on their net radiative effects. These clouds all have tops colder than 245 K, and the optical depth ranges are thin (0-4), anvil (4-32) and thick (32-64). Defined in this manner, thin clouds have a TOA warming effect, and anvil and thick cloud a TOA cooling effect. We reserve a separate category for thick cloud because these likely represent deep convective cores.

The average-sky cloud forcing histograms, which hereafter we will refer to as net CRF, are presented in Fig. 5 for the WP, CP, and EP. The net CRF in the EP is much more negative (-51 W m\(^{-2}\)) than the CP (-22 W m\(^{-2}\)) or WP (-16 W m\(^{-2}\)) CRF, partly due to the larger population of low, thick cloud there. Also, the WP (and CP, to a slightly lesser extent) contains much more high, thin cloud, which contributes to a smaller negative net CRF.

Fig. 5 also gives the net CRF value for only high cloud in each of the three regions, including only clouds colder than 245 K. The WP and CP have similar high net CRF, at -7.5 and -8.4 W m\(^{-2}\), respectively. This contrasts with the EP, which has a high CRF of -18.2 W m\(^{-2}\). The less negative high cloud forcing in the WP stems from differences in thin cloud amount, which will be more clearly demonstrated in the next section. Differences in anvil and thick cloud net CRF will also be addressed later on.

5. High Cloud Amount Versus Convective Strength

Simple energy balance indicates that net heating by precipitation in convecting regions is balanced by radiative cooling in adjacent subsidence regions. Deep cumulonimbus clouds carry water vapor from the shallow cloud layer to the upper troposphere, and as they do, the water
vapor condenses. It is these cumulonimbus towers that precipitate most heavily, and thus it should follow that rainrate is a good proxy for convection strength. We have composited high cloud with rainrate, with the hope of revealing possible regional differences in characteristics of convection along the ITCZ. By compositing with rainrate, we effectively remove convection strength from the equation, such that differences in structure for a given convective regime can be compared. In Fig. 6, thin, anvil and high thick cloud amount are presented as a function of rainrate, averaged over the two-year period of September 2003 through August 2005. Each point represents the following rainrate percentiles ranges: 1) 2.5th-10th, 2) 10th-25th, 3) 25th-50th, 4) 50th-75th, 5) 75th-90th, and 6) 90th-97.5th. Each error bar represents the observed value ±3 times the sampling error. The sampling error is given by

\[
\frac{\sigma}{\sqrt{N}}
\]  

(2)

where \( \sigma \) is the standard deviation of all 1° by 1° boxes (three-day aggregates) within each rainrate percentile regime that have nonzero precipitation rates and at least 200 good data points, and \( N \) the number of 1° by 1° boxes (three-day aggregates) that fall within each rainrate regime. Sampling error bars encompassing ±3 times the sampling error represent the 99% confidence interval, assuming normally distributed data. It is important to note that the 1° by 1° boxes are assumed to be independent of one another, which may not necessarily be the case, such that the error bars might be underestimates.

On the top panel of Fig 6, we see that for a given rainrate, thin cloud is more than twice as abundant in the WP as in the EP. Also, thin cloud is nearly independent of rainrate. The observed thin cloud fraction decreases slightly with rainrate as anvil and thick cloud increases with rainrate, and this is because as there is more opaque cloud, thin cloud is not as detectible by MODIS, even if it is present.
Anvil cloud is more abundant in the WP than the CP or EP for a given rainrate, though the differences are less dramatic than for thin cloud. Finally, thick cloud in the WP, CP, and EP increases in a very similar manner with rainrate, and it is difficult to distinguish between the three regions. This suggests that thick cloud amount is a very good proxy for rainrate, which could be quite practical for remote sensing purposes, as using OLR alone as a proxy for rainrate can give fallacious results (Berg et al. 2002).

The fact that combined anvil plus thin cloud combined is much more abundant in the WP per unit of precipitation, despite thick cloud being the same, implies important structural differences in convection in the WP versus the EP. If we understand convective cores to be manifested as thick, high cloud that then spreads and thins out with time, then the thick cloud in the WP that spreads out into thinner cloud shields is sustained over larger geographical regions. In a later section, we will show that relative humidity is higher in the WP in the upper troposphere, and perhaps the moister upper troposphere there slows the evaporation of the ice compared to the EP. We cannot prove this latter statement, however, as it is probable that convection contributes to the observed upper-level relative humidity profiles (Udelhofen and Hartmann 1995, Sassi 1995). The EP ITCZ is considerably narrower than the WP ITCZ, so that the EP is surrounded by a much drier upper-level environment. The EP is thus more readily influenced by the surrounding dry regions, which could be less conducive to maintenance of anvil and thin high cloud. Entrainment of non-ITCZ air into the EP in the upper troposphere could partly explain the lower anvil and thin cloud amount. At present, we cannot compute accurate moisture budgets for these regions, however.

We also examine the TOA radiative impacts of thin, anvil, and thick cloud by compositing them with rainrate. Fig 7 shows that thin cloud increases TOA radiation, and anvil
and thick cloud reduce the TOA energy budget. For a given rainrate, the CRF of thin high cloud is approximately 10 W m\(^{-2}\) higher in the WP versus the EP. With increasing rainrate, the CRF of anvil and thick cloud becomes more negative, as these cloud types become more abundant. The anvil and thick net CRF in the WP and EP are nearly indistinguishable from one another, however, despite the presence of more anvil cloud in the WP. This is because even though the SW effect is \textit{stronger} for anvil cloud (because the anvil cloud fraction is greater) for a given rainrate in the WP (not shown), the LW effect is also \textit{stronger} (also not shown), since anvil cloud is slightly colder in the WP.

Finally, the bottom right panel of Fig. 7 shows that the total high CRF is considerably more negative for the EP for a given rainrate, which we can safely say is mostly due to much less thin, high cloud there, since the CRF due to the other high cloud types (anvil and thick) is very similar in the WP and EP for a given rainrate. Thus, the structural differences in convection in the WP and EP also have significant implications for the energy balance at the top of the atmosphere.

6. High Cloud Top Temperature Versus Rainrate

We now present thin, anvil, and thick cloud top temperature versus rainrate in Fig. 8. It is clear that all high cloud types in the WP and CP are systematically about 5 K colder than the EP, for a given rainrate. Once again, for quality assurance purposes, these data only include each three-day 1° by 1° box for which at least the cloud fraction is greater than zero, at least 200 good 5-km data points are present, and the rainrate is nonzero. For all rainrates during the September 2003 through August 2005 period, the median cloud top temperatures of thin clouds are WP=217.4 K, CP=217.6 K, and EP=222.8 K. For anvil cloud, the respective temperatures are 219.4
K, 220.8 K, and 224.3 K, and for thick cloud 212.1 K, 214.3 K, and 218.5 K. Thus, WP thin and anvil clouds are approximately 5 K colder than in the EP, and thick cloud about 6 K colder.

A few points can be made about these observations. First, as expected, thick convective cores penetrate highest in the atmosphere, and anvil clouds detrain from these cores at a lower level. Thin cloud is only slightly colder than the anvil cloud, which seems to lend credence to a notion that it is related to the anvil cloud. It is important to note, however, that MODIS does not see tropopause cirrus at all, as it is too optically thin. One can view the evolution of convection such that anvil cloud detrains from core clouds, and over time, thin cloud is the residual of the anvil cloud as it spreads and thins out away from the convective cores, although some high thin cloud can be generated by waves (e.g. Boehm and Verlinde 2000). It seems that MODIS can detect thin cloud at realistic temperatures, compared to satellite estimates based on thermal imaging alone, for which the temperature depends on emissivity. The CO$_2$ slicing method utilized by MODIS, on the other hand, does not require an emissivity estimate.

It is noteworthy that anvil and thin cloud top temperature seem to show no clear relationship with rainrate, whereas thick cloud (save the lowest rainrates in the WP) seems to get colder with increasing rainrate, at least in the EP. This perhaps is consistent with the notion that the thick cloud is convectively driven, and since rainrate is a proxy for convection strength, we might expect the deepest convection be more sensitive to the SST.

**7. Clear-Sky Divergence Calculations in the Upper Troposphere**

We wish to better understand cloud top temperature differences in the WP and EP, and to ask how the cloud top temperature is determined. We explore both the 'PUSH' and 'PULL' concepts; the former suggests that the altitude of anvil and thick clouds is driven by buoyancy, while the latter argues for clear-sky radiation being the key factor. In this section, we focus on
data use to explore the PULL concept. Calculation of clear-sky convergence profiles requires moisture and temperature profiles, that are then used to calculate clear-sky cooling rates and vertical velocity (ω) profiles.

The Microwave Limb Sounder (MLS), aboard the Aura satellite, uses the 190 GHz channel to measure upper-level water vapor mixing ratios. There are 22 levels between 316 hPa and 0.1 hPa, with measurements made in the upper tropospheric region of interest at 316 hPa, 215 hPa, 147 hPa, and 100 hPa (Livesey et al. 2005). This corresponds to a vertical resolution range of 2.7-3.0 km. This resolution in the upper troposphere with MLS is superior compared to more conventional measurements of upper-level relative humidity from radiosondes, whose spatial resolution is spotty, and satellites, whose vertical resolution is coarse (Sassi et al. 2001). Vertical resolution in the upper troposphere is important for moisture for clear-sky divergence calculations. Also, MLS can measure upper tropospheric moisture in clear and cloudy regions, as opposed to nadir satellite measurements using thermal infrared, which are limited to clear skies. Data from MLS aboard the Aura satellite are available from September 2004 onward.

Vertical temperature profiles are available from Global Positioning System (GPS) data, with 0.5 K precision. Though GPS data are available well before September 2004, we start from September 2004 to overlap with the Aqua data. This is a navigation system which contains 24 satellites. Major advantages of GPS include long-term stability, good vertical resolution, all-weather capability, and global coverage. Atmospheric refractivity fields are utilized, allowing conversion to profiles of temperature and pressure (Schmidt et al. 2004). The largest temperature errors occur near the surface and at 8 km, and are 0.5 K and 0.2 K, respectively. Errors fall to 0.1 K in the majority of the stratosphere (Kursinski et al. 1997).
The same radiative transfer model used for CRF calculations is used to compute atmospheric profiles of clear-sky radiative cooling rates, which require temperatures and mixing ratios as inputs. GPS temperatures are interpolated onto the 1000 pressure levels in the model; the layers between each level contain approximately equal mass. Between 0.1 hPa and 147 hPa, MLS mixing ratios are interpolated onto the model pressure levels. For mixing ratios between 147 hPa and 316 hPa, a polynomial fit of order three is used to interpolate data onto the model pressure levels. Between 316 hPa and 375 hPa, a linear fit is made, and below this, an idealized tropical profile (with specific humidity) is used, from McClatchey et al. (1971).

Once the radiative cooling rates are calculated, they are used to calculate the diabatic omega (\(\omega\)) profiles from the following relationship, which is derived from the thermodynamic energy equation (Holton 1992):

\[
\omega = \frac{-Q(p)pc_p}{R_D T} \left[1 - \frac{pc_p \partial T}{R_D T \partial p} \right]^{-1}
\]  

In (3), Q(p) is the radiative cooling rate as a function of pressure, \(c_p\) is the specific heat at constant pressure \([=1004 \text{ J K}^{-1} \text{ kg}^{-1}]\), \(R_D\) the dry air gas constant \([=287 \text{ J K}^{-1} \text{ kg}^{-1}]\), T the absolute temperature, and \(\omega\) the vertical velocity in pressure coordinates.

Next, the horizontal divergence is calculated from the continuity equation, which is given by

\[
\text{Div}_H = -\frac{\partial \omega}{\partial p}
\]  

In (4), \(\text{Div}_H=\partial u/\partial x+\partial v/\partial y\). Because of the derivative, the estimated divergence is quite noisy, and a smoothed divergence profile is computed for each region by averaging into five-degree temperature bins corresponding to the temperature bins used for the MODIS data. Vertical
profiles of relative humidity, cooling rates, $\omega$-profiles, and smoothed divergence profiles are discussed in more detail in the next section.

8. Relative humidity, cooling rate, $\omega$, and clear-sky convergence profiles

Composite relative humidity profiles from September 2004 through July 2005 for the WP, CP, and EP are shown in Fig. 9. The focus here is in the temperature range where anvil clouds detrain, which occurs between about 215 K and 225 K. In the EP and WP profiles a secondary peak in relative humidity occurs at about these temperatures. The peak in relative humidity in the WP is larger in magnitude and peaks at a lower temperature than the EP. These differences in relative humidity peak temperature and amplitude are interesting, and might be related to the fact that there are simply more anvil and thin cloud in the WP compared to the EP.

Clear-sky radiative cooling rate profiles are presented in Fig. 10, and show that below 250 hPa, cooling rates are roughly constant at around 2 K day$^{-1}$. Above this layer, clear-sky radiative cooling rates drop off sharply. Larson and Hartmann (2002) argue that this is because of the Clausius-Clapeyron relationship, which states that the saturation vapor pressure depends only on temperature. Clear-sky emission from the upper troposphere comes from the rotational lines of water vapor, and depends strongly on vapor pressure. The sharp decreases of clear-sky radiative cooling rate profiles above 250 hPa occur because of low absolute temperatures there and consequently low vapor pressures. The large decrease in clear-sky radiative cooling rates corresponds to a mass convergence at this temperature, and this balances divergence from convective regions at the same temperature. It is this circulation that induces detrainment of the majority of anvil cloud.

Profiles of clear-sky $\omega$ are shown in Fig. 11, and indicate fairly strong subsidence below about 250 hPa (on the order of 40 mb/day), and decreasing sinking motion above this level. The
Q=0 level, which is the level at which clear-sky radiative cooling is zero, is just below the tropopause (which is around 95 hPa in all three regions), and well above the maxima in clear-sky subsidence.

Smoothed clear-sky convergence profiles are shown in Fig. 12. Clear peaks appear in the upper-troposphere in the WP and EP, roughly corresponding to the peaks in upper-level relative humidity. To quantify the differences in the WP and EP, a convergence-weighted temperature (T_C) is calculated from the following relationship:

$T_C = \frac{1}{C\Delta T} \int_{190K}^{240K} C(T)TdT$ \hspace{1cm} (5)

where C(T) is the smoothed convergence, \( \overline{C} \) is the mean convergence in the layer from 240 K to 190K, and \( \Delta T = 50 \) K. \( T_C \) provides a quantitative measure of the temperature at the peak of convergence, and is 216.6 K in the WP and 220.2 K in the EP for the period of September 2004 through July 2005. As we shall see in a later section, these differences in clear-sky convergence correspond rather well with differences in anvil cloud top temperature measured by MODIS.

In an attempt to quantify why \( T_C \) is about 3.6 K warmer in the WP versus the EP, we explore the relative importance of specific humidity and temperature profiles. We do this by pairing the WP and EP temperature and specific humidity profiles in all possible combinations and computing the corresponding \( T_C \). Table 1 indicates that the specific humidity difference is more important than the temperature, but the temperature is also a factor. Fig. 4 shows that the WP has a larger lapse rate in the region above 12 km, or colder than about 225 K. An idealized tropical profile, from McClatchey al. (1971), was used below the MLS levels. While reanalysis data below the MLS region would perhaps be preferred, sensitivity studies were performed
(figures not shown), and suggested that moisture further down in the troposphere does not matter very much at all compared to moisture closer to $T_C$.

9. Testing the PUSH Concept

The notion behind the PUSH concept is that the surface drives convection, by assuming that parcels of air, so long as they retain positive buoyancy, ascend undiluted until they become negatively buoyant in the upper troposphere (Riehl and Malkus 1958, Riehl and Simpson 1979). The level of neutral buoyancy in the upper troposphere is reached at the temperature aloft in which environmental $\theta_E$ is the same as $\theta_E$ at the LCL. We refer to this intersection temperature as $T_{INT}$. We calculate $T_{INT}$ first by calculating $\theta_E$ profiles for the WP, CP, and EP, and then simply determining the intersecting point in the upper troposphere. Given this procedure, parcels in the WP have the potential to ascend higher, as $\theta_E$ at the LCL is 365.8 K, versus 358.4 K in the EP (these are calculated assuming that the LCL temperatures are 6K lower than the SSTs in each of the regions). For the same period (September 2004 through July 2005), $T_{INT}$ is 192.9 K in the WP, 194.9 K in the CP, and 199.8 K in the EP. These temperatures are significantly lower than $T_{ANVIL}$ (median anvil cloud top temperature), by 20 K or more, and even $T_{THICK}$ (median thick cloud top temperature). Therefore, the surface air in both the EP and WP has enough energy to rise undiluted to well above the level where the anvil cloud is concentrated. One must assume a particular entrainment spectrum for plumes in order to explain the existence of the anvils. The clear-sky radiative cooling profile can explain the temperature of the anvil clouds much more fundamentally.

10. Comparing the Push and Pull Concepts

It was discussed previously that clear-sky convergence profiles support a circulation connecting clear and adjacent convective regions. The peak of the clear-sky convergence should
therefore correspond to the detrainment level of anvil cloud. To investigate this, we examine a subset of data, from September 2004 through July 2005, during which we have GPS, MLS, and MODIS data. In Fig. 13, we present the clear-sky convergence profiles along with the anvil cloud fraction profiles for the WP and EP, both as functions of temperature. The peak in upper-level convergence lines up well with the peak in anvil cloud amount in all three regions (only the WP and EP are shown for clarity). The differences in $T_{\text{ANVIL}}$ in the WP and EP correspond well to those of the clear-sky convergence peak. Thus, vertical profiles of convergence and observed anvil cloud fractions are related to one another.

In order to further quantify whether or not $T_C$ is actually a good predictor of $T_{\text{ANVIL}}$, we subdivide the data into seasonal datasets, so as to increase the number of samples with which to evaluate the statistical validity of the relationships between $T_C$ and $T_{\text{ANVIL}}$. We present $T_{\text{ANVIL}}$ versus $T_C$ in Fig. 14, as well as the one-to-one line of these two variables. Principal component analysis is performed in this plot and subsequent ones to calculate the slopes, which makes no assumptions of dependent or independent variables. The slope of the best linear fit is $1.06 \pm 0.41$, which means that a one-to-one relationship between $T_{\text{ANVIL}}$ and $T_C$ cannot be ruled out. $T_{\text{ANVIL}}$ is only about three to four degrees Kelvin warmer than $T_C$. Fig. 14 shows that the peak in clear-sky convergence corresponds well to $T_{\text{ANVIL}}$, such that the clear-sky radiative profile seems to drive anvil cloud detrainment!

The PULL method certainly seems to be able to predict the level of anvil cloud detrainment. Can the PUSH method also be used to predict the level of convective detrainment? We address this first by plotting $T_{\text{ANVIL}}$ versus $T_{\text{INT}}$ for the seasonal data (Fig. 15), and immediately see, as suggested earlier, that $T_{\text{INT}}$ is much lower (by 20 K or more) than $T_{\text{ANVIL}}$. 
The slope of the best linear fit line is $0.59 \pm 0.41$, so that it seems that $T_{\text{INT}}$ does not explain where anvil clouds are detraining.

But, does $T_{\text{INT}}$ have any bearing on how high the thickest clouds get? We might expect more of a linear relationship between the thickest high clouds and $\Theta_E$ near the surface, since they represent the tallest convective cores, which are least affected by detrainment. Fig. 16 shows both $T_{\text{THICK}}$ versus $T_{\text{INT}}$ and the 90th percentile of the coldest thick clouds from MODIS ($T_{\text{THICK,90}}$) versus $T_{\text{INT}}$. We immediately see that $T_{\text{INT}}$ is significantly colder than $T_{\text{THICK}}$ by approximately 10-15 K, but that the slope of the best linear fit is $0.96 \pm 0.41$, so a one-to-one relationship cannot be ruled out. The relationship between $T_{\text{INT}}$ and $T_{\text{THICK,90}}$ is more convincing, especially since the 90th percentile of the thick clouds often makes it to $T_{\text{INT}}$. The slope of the line is $0.69 \pm 0.38$. Also, the fact that $T_{\text{INT}}$ and $T_{\text{THICK,90}}$ are close to one another does seem to suggest that there is a relationship between the two. The low-level $\theta_E$ thus does seem to constrain the coldest level of the convective cores, as we would expect, but the anvil cloud temperature is better predicted by the clear-sky convergence.

11. Conclusions

We have examined structural characteristics of tropical convection in the ITCZ of the North Pacific. We have shown three modes of clouds in the WP, CP, and EP, namely trade cumulus clouds, mid-level congestus clouds, and deep convective cores and the associated detrained anvil and thin clouds. Low cloud is more prevalent in the EP, where the domain chosen is surrounded by low SST and abundant stratocumulus cloud. Congestus cloud is also somewhat more prevalent in the EP. This may be a result of the fact that convection in the EP is more strongly driven by surface convergence (Back and Bretherton 2006).
For a given rainrate, high thin cloud is nearly twice as abundant in the WP as compared to the EP, and this makes the TOA radiative forcing due to this cloud considerably higher in the WP, on the order of 10 W m\(^{-2}\). Anvil cloud fraction for a given rainrate is also somewhat higher in the WP compared to the EP, though because the anvil cloud is colder in the WP, the net radiative forcing of anvil cloud is about the same in the two regions. Thick cloud amount has the same relationship with rainrate in the WP, CP, and EP, making it a very good proxy for rainrate.

The differences in high cloud radiative forcing in the WP and EP stem primarily from differences in high thin cloud amount. Thin and anvil clouds are systematically about 5 K colder in the WP versus the EP, and thick clouds about 6 K colder. Thick clouds are the coldest, and anvil and thin clouds are observed at similar temperatures. It seems that anvil clouds detrain from convective cores, and over time, thin cloud is the residual of the anvil cloud. We speculate that higher amounts of thin and anvil cloud in the WP are due to the broad structure of the ITCZ there, whereas the domain chosen for the EP is surrounded by subsidence regions, whose free-tropospheric air is drier, and may entrain somewhat into the EP convective zones.

We also examine how cloud top temperature is determined by examining the PUSH and PULL concepts. To examine the PULL concept, which states that convective detrainment is driven by clear-sky convergence, we first examine upper tropospheric relative humidity. It is higher in the WP, and peaks at a lower temperature than the EP. The relative humidity structure seems to be consistent with the abundance of anvil and thin cloud, with greater humidity associated with more high cloud. Also, seasonal data suggest that the temperature weighted by clear-sky convergence is a good predictor of \(T_{\text{ANVIL}}\). This suggests that the level of maximum clear-sky convergence in the upper troposphere *drives* the level of anvil cloud detrainment.
In examining the PUSH method, on the other hand, the temperature where near-surface air achieves neutral buoyancy is much colder than $T_{\text{ANVIL}}$, and $\Theta_E$ does not seem to control $T_{\text{ANVIL}}$. The thickest, coldest clouds (the 90th percentile of the thick clouds) can make it to $T_{\text{INT}}$, so near surface $\Theta_E$ might be controlling the altitude of the coldest convective cores.

Acknowledgements. This work was supported by NASA Grant NNG05GA19G.
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Fig. 1. Map of Composite SSTs over the tropical Pacific for September 2003-August 2005, indicating the three regions selected for study.
Fig. 2. West Pacific shortwave (top), longwave (middle), and net (bottom) cloud radiative forcing, assuming 100% cloud percent in each T-τ bin.
Fig. 3. T-τ histograms of cloud percent for September 2003-August 2005, for WP, CP, and EP (top to bottom).
Fig. 4. GPS temperature profiles (top) and WP-EP temperature differences (bottom) for September 2004-July 2005.
Fig. 5. Cloud radiative forcing contour plots binned by T-τ for WP, CP, and EP (top to bottom) for September 2003-August 2005.
Fig. 6. Thin, anvil, and thick cloud fraction versus rainrate for September 2003-August 2005.
Fig. 7. High cloud cloud radiative forcing versus rainrate for the same period as Fig. 6.
Fig. 8. Cloud top temperature versus rainrate for thin (top), anvil (center), and thick (bottom) high cloud for September 2003-August 2005.
Fig. 9. September 2004-July 2005 relative humidity with respect to ice for WP, CP, and EP. Also shown are MLS levels.
Fig. 10. Calculated clear-sky radiative warming rates for WP, CP, and EP.
Fig. 11 Vertical velocity in pressure coordinates ($\omega$) necessary to balance clear-sky radiative cooling for September 2004-July 2005.
Fig. 13. Clear-sky convergence and anvil cloud fraction for WP (top) and EP (bottom) for September 2004-July 2005.
Fig. 14. Seasonal median anvil cloud top temperature ($T_{ANVIL}$) versus seasonal convergence-weighted temperature ($T_C$).
Fig. 15. Seasonal median anvil cloud top temperature ($T_{\text{ANVIL}}$) versus seasonal intersection temperature ($T_{\text{INT}}$ is the temperature in the upper troposphere where $\Theta_E$ is the same as $\Theta_E$ at the LCL).
Fig. 16. Seasonal thick cloud temperature ($T_{THICK}$) and 90th percentile of $T_{THICK}$ versus seasonal intersection temperature ($T_{INT}$ is the temperature in the upper troposphere where $\Theta_E$ is the same as $\Theta_E$ at the LCL).
Table 1. Convergence-weighted temperature ($T_C$) for WP and EP specific humidity and temperature profiles.

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<tr>
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<th>WP Specific Humidity</th>
<th>EP Specific Humidity</th>
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<tr>
<td>WP Temperature Profile</td>
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<td>219.2 K</td>
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<tr>
<td>EP Temperature Profile</td>
<td>218.1 K</td>
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