A Multi-Sensor Perspective on the Tropical Interannual Variability of Humidity and Clouds

A dissertation submitted in partial satisfaction
of the requirements for the degree
Doctor of Philosophy in Atmospheric and Oceanic Sciences

by

Calvin K. Liang

2011
The dissertation of Calvin K. Liang is approved.

________________________
Annmarie Eldering

________________________
Alex Hall

________________________
Jochen Stutz

________________________
Charlie S. Zender

________________________
Kuo-Nan Liou, Committee Chair

University of California, Los Angeles
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“I am my Beloved’s and my Beloved is mine”

—Song of Solomon 6:3

For my Beloved.
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Vita

1978 Born, Monterey Park, California, USA.

1996 Home Savings and Loan Teller, Rosemead, California

2000 A.B. (Physics), Occidental College, Los Angeles, California.

2004 Adjunct Instructor of Physics (Mechanics), Occidental College, Los Angeles, California.

2005 Adjunct Instructor of Physics (Light), Occidental College, Los Angeles, California.

2000-11 Member of the Technical Staff, Northrop Grumman Aerospace Systems (formerly TRW), Redondo Beach, California.

2006 Teaching Assistant (AOS 1), Department of Atmospheric and Oceanic Sciences, UCLA, Los Angeles, California.

2007 M.S. (Atmospheric and Oceanic Sciences), UCLA.

2007-11 Graduate Student Researcher, Department of Atmospheric and Oceanic Sciences, Joint Institute for Regional Earth System Science and Engineering, UCLA.


A Multi-Sensor Perspective on the Interannual Variability of Tropical Temperature, Water Vapor, and Clouds (December, 2010). American Geophysical Union Fall Meeting, San Francisco, California

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ABSTRACT OF THE DISSERTATION

A Multi-Sensor Perspective on the Tropical Interannual Variability of Humidity and Clouds

by

Calvin K. Liang

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Professor Kuo-Nan Liou, Chair

The distribution of water vapor (H₂O) in upper troposphere/lower stratosphere (UTLS) has significant impacts on the tropical radiative structure. Much analyses has been done on quantifying the various processes that govern the UTLS H₂O budget, particularly in the tropical tropopause layer (TTL) (~150-70 hPa [FDD09]). However, this task has been difficult due to the lack of high spatial and temporal measurements of H₂O. Fortunately, a new generation of passive and active sensing instruments, flying on the NASA “A-Train” constellation of satellites, provide a unprecedented three dimensional perspective of the distribution of humidity and clouds. These include soundings from the Aqua Atmospheric Infrared Sounder (AIRS) and Aura Microwave Limb Sounder (MLS), and cloud profiles from the CloudSat and Cloud-Aerosol Lidar and Infrared Pathfinder Satellite Observations (CALIPSO). From these measurements, we develop a new integrated dataset of H₂O, temperature, and cloud properties to characterize the tropical distribution of humidity, clouds, and radiative heating.

We first characterize the sensitivity of both AIRS and MLS to UTLS H₂O. We find that the instruments overlap in sensitivity over a half temperature scale.
height layer making it possible to fuse their soundings into a single profile that spans the entire troposphere and stratosphere. The method we develop to combine the H$_2$O soundings produce profiles that are not only smooth functions with altitude but preserve the atmospheric state as interpreted by both the AIRS and MLS instruments.

From these profiles, we quantify the tropical interannual variability of temperature and H$_2$O for the period when both AIRS and MLS are co-located. From this analysis we confirm that the El Niño Southern Oscillation (ENSO) and the quasi-biennial oscillation (QBO) both significantly impact the tropopause region temperature and H$_2$O structure. Furthermore, we find differing regional impacts on temperature depending on the relative phase of the ENSO and QBO. When these tropical modes are in phase the tropical western Pacific (TWP) experiences temperature anomaly enhancements while the tropical central Pacific (TCP) anomalies are reduced. When the ENSO and QBO fall out of phase, the additive behavior of the temperature anomalies migrate to the TCP with the TWP experiencing anomaly reductions. It is hypothesized that this migration of anomaly enhancement and cancellation are due to changes in the Walker Circulation, i.e. migration of convection between the TCP and TWP, resulting from oscillations between El Niño and La Niña years. During La Niña (El Niño) years the tropical zonal dehydration is enhanced (reduced) especially when the ENSO and QBO are in phase [ZGZ04]. This is primarily due to the temperature fluctuations in the TWP tropopause region [HG01, ZGZ04]. However, when the ENSO and QBO are out of phase the TCP H$_2$O distribution becomes a significant contributor to the zonal mean H$_2$O distribution. This suggests the assumption that dehydration in the TWP mainly governs the tropical zonal H$_2$O distribution may need to be reexamined.
Furthermore, we characterize the interannual impacts of the ENSO on the tropical cloud distribution. The CloudSat and CALIPSO measurements affirm the ENSO impact on the Walker Circulation. However, we also show that the ENSO causes a migration between the TCP and TWP of a majority of tropical cloud types with most clouds below $\sim 15$ km following the vertical $\text{H}_2\text{O}$ structure and the thin cirrus cloud above $\sim 15$ km following an eastward tilt in temperature due to eastward propagating Kelvin waves induced by tropical convective heating. This not only confirms the water vapor and temperature effects of ENSO on tropical clouds but bolsters the hypothesis that temperature variance has a dominant influence on the distribution of tropical thin clouds in the TTL. This is important given the greater abundance of thin cirrus clouds, up to an order of magnitude, relative to other clouds. We also find evidence that the temperature variations imposed by the zonally symmetric QBO also impacts the distribution of thin cirrus amount. We show, from radiative transfer calculations, that these thin clouds have the largest impact on the tropical TTL radiative balance, specifically the location of the level of zero radiative heating (LZRHR) [CLP05, GFM04].
CHAPTER 1

Introduction

“Let’s Get It Started” – Black Eyed Peas

It has been known for sometime that air enters the stratosphere via the tropical tropopause [Bre49]. Although increases in tropospheric water vapor (H$_2$O) are expected in a warming climate, perturbations to its amount in the upper-troposphere (UT) produces the largest radiative changes relative to other regions of the atmosphere [HS00]. Lower stratospheric (LS) H$_2$O also can have a significant impact on climate. The recent reduction in LS H$_2$O (mean difference taken between 2001-2005 and 1996-2000) resulted in a radiative forcing of -0.098 W/m$^2$ which offset part of the +0.26 W/m$^2$ forcing from the increase in CO$_2$ from 1996-2005 [SRP10]. Furthermore, changes to stratospheric H$_2$O may have consequences on future stratospheric ozone concentrations [KHA99]. Changes in H$_2$O will also affect the frequency of cirrus occurrence, therefore, also impacting the radiative balance of the UTLS (∼300-70 hPa) [GFM04, CLP05]. Thus, quantifying the processes that govern the distribution of UTLS H$_2$O is crucial.

The UTLS H$_2$O budget is largely determined by contributions from localized convective moistening [Sod04, HS07], dehydration [FBH05, FH05], horizontal advection through cold trap regions, precipitating thin cirrus [HG01], and convective mixing [SD01, SD03]. Furthermore, interannual variability modes such
as the El-Niño Southern Oscillation (ENSO) [Phi90, WRM98] and the Quasi-Biennial Oscillation (QBO) [Ree65b, Ree65a, Ham84, Dun85], also affect the UTLS temperature, and thus, H₂O distribution. [GRM01] showed that the ENSO modulates tropopause temperatures in the TCP, making them colder (warmer) during El Niño (La Niña) periods. [ZGZ01b] showed the same results for the tropical cold point tropopause (CPT). The QBO manifests itself as descending temperature anomalies [AHL87] in thermal wind balance with the descending westerly/easterly wind anomalies, which are a result of the interaction between the zonal mean wind and the upward propagating equatorial waves (from tropical convection) in the stratosphere [BGD01]. Work by [FBH05] and [FH05] showed that UTLS H₂O concentrations can be reconstructed tracing a parcels history back to the coldest temperature it experiences–this is termed the Lagrangian Cold Point (LCP)–and that the interannual modes of the ENSO and QBO both impact the LCP.

Another element deeply connected to the UTLS temperature and H₂O structure is the tropical high cloud distribution. Equatorial tropical heating leads to a deep convective influence on UTLS temperature variance through excitation of Kelvin and Rossby wave propagation [Gil80]. The convective wave modes that impact the temperature distribution, in turn, impact the tropical cirrus distribution, particularly tropical thin clouds [VWF10]. Recent work has also showed that tropical deep convection interacts with the large scale circulation in the Madden Julian Oscillation (MJO) with a time-scale of ∼30-60 days. It was shown that deep convective heating provides the large-scale waves that pre-humidifies the troposphere which organizes more deep convective cells into mesoscale convective systems (MCS) [TR10]. These MSCs are the cloud systems that provide the dominant source of latent and radiative heating in the tropical troposphere (lower and upper). In effect, these waves provide a positive feedback, via the
convective systems, to amplify the impacts of these waves. The deep convection that impacts the UTLS temperature structure, makes the UTLS colder above the region of convective heating [SD01, HN07]. These are the regions where tropical tropopause layer (TTL, layer between 150-70 hPa [FDD09]) cirrus persistently form [CAT04].

The tropical structure observed on intraseasonal time scales (∼30-60 days) also appear on interannual time scales (∼1-5 years) [FDD09]. The cloud structure associated with ENSO is mainly observed in the regional changes of cloudiness between the tropical western Pacific (TWP) and tropical central Pacific (TCP) associated with shifts in the Walker-Circulation [Phi90, GRM01]. Once again, the waves associated with deep convection impact the TTL temperature structure, thus impacting cloud amount. There is also evidence that the phase of the QBO can impact tropical convection as well [CHM98, CMH02].

Although significant progress has been made in understanding the processes that govern the interannual variations of UTLS temperature and H₂O, quantifying the effects of the ENSO and QBO on these quantities has been difficult due to a lack of measurements that span the troposphere and stratosphere. Previous satellite data records had limits in their vertical sensitivity, resolution, and spatial extent. For example, the GOES Visible Infrared Spin Scan Radiometer (VISSR) Atmospheric Sounder (VAS) was only able to retrieve H₂O in a single column mean layer over a depth of a few hundred millibars within the UT and nowhere below [SB93]. The Halogen Occultation Experiment (HALOE) [RTG93] and the Microwave Limb Sounder (MLS) [WCC99] on the Upper Atmosphere Research Satellite (UARS) used limb measurements to provide global coverage of H₂O measurements primarily in the LS. Temperature and H₂O measurements that span the entire troposphere and stratosphere are limited to sparsely located
in-situ instruments, making it difficult to explore the interaction between the ENSO and QBO.

The AIRS and MLS instruments, on board the A-Train ("A" is for "Afternoon") constellation, provide high quality UTLS H\textsubscript{2}O measurements, however, AIRS primarily has sensitivity to H\textsubscript{2}O consistent with concentrations from the boundary layer up to the UT. The lower threshold of AIRS sensitivity to H\textsubscript{2}O was derived empirically, in [GWF04], to be \sim 10 ppmv based on comparisons with in-situ measurements; [FRW08] found the lower threshold to be between 20–30 ppmv. MLS, however, has sensitivity in the stratosphere down to the UT with sensitivity to values between \sim 0.1–1 ppmv, but with an upper sensitivity limit of \sim 500 ppmv [RLB07]. Studies by [RLB07] and [FRW08] attempted to quantify the sensitivity of AIRS and MLS H\textsubscript{2}O by direct comparison of retrieved values and found that AIRS and MLS agreed best between \sim 250–260 hPa.

The vertical cloud distribution has also been difficult to quantify due to limitations of sounding instruments not being able to easily distinguish cloud layers. The most comprehensive and extensive cloud record up to date comes from the International Satellite Cloud Climatology Project (ISCCP) [RS99], but does not provide detailed information on the vertical structure of clouds. The vertical structure is important because of the radiative impacts of high tropical clouds in the UTLS [GFM04, CLP05, CLF06]. Particularly, the vertical structure determines the level of zero radiative heating (LZRH), a quantity that describes the location a parcel of air must be lifted, in the absence of convection, in order for the background environment to lift it into the stratosphere.

In order to quantify the interannual distribution of clouds and their radiative impacts through the entire tropical atmosphere, we utilize the newly available CloudSat [SVB02] and Cloud-Aerosol Lidar and Infrared Pathfinder Satellite
Observations (CALIPSO) [WPM03] data which actively profile the atmosphere, providing a high vertical resolution information on the layer structure of clouds. Both of these datasets have provided the ability to jointly identify the relationship between temperature, H$_2$O, and clouds [KLE08, KGF09]. [YFH10] also used these cloud profiles in concert with a radiative transfer model (RTM) to characterize the radiative heating associated with thin cirrus in the TTL.

For this work (published in [LEI10]), we construct a combined temperature and H$_2$O that provides a high fidelity and vertically resolved dataset to quantify the climatological distribution of these quantities with a particular focus on their interannual variations. The newly available AIRS (in the version 5 dataset) and MLS (offline calculation) averaging kernels (AK’s) are analyzed to quantify where both instruments are most, and least, sensitive to H$_2$O in the UTLS. Previous work by [MB08] computed AIRS AK’s for selected sites and approximated the instrument resolution for temperature and H$_2$O. We use the AK’s for an entire year of AIRS retrievals to quantify the instrument’s sensitivity to UTLS H$_2$O. We use this information as a means to merge AIRS and MLS H$_2$O retrievals into a single profile which spans the troposphere and stratosphere.

We then develop a tropical time record of H$_2$O, temperature, and relative humidity ($RH$), for pressures between 1013 hPa and 10 hPa, from a newly combined Atmospheric Infrared Sounder (AIRS) [ACG03, SBB06, DBG06, FRW08, CPA06] and Microwave Limb Sounder (MLS) [WFH06, RLB07] dataset [LEI10] for the majority A-Train record (Aug, 2004-Mar, 2010) when both instruments have co-located measurements (see Section 3.2 for details). We use these profiles to characterize the tropical distribution of temperature and H$_2$O, specifically quantifying the joint impacts of the ENSO and QBO around the tropopause region (published in [LEG11]). The Tropical Rainfall Measuring Mission (TRMM) pre-
cipitation dataset [IKM00] is incorporated into this study as a proxy for deep convection.

Furthermore, we construct a global dataset of cloud frequency of occurrence (CFO) to identify the impacts of ENSO and the QBO on the entire cloud distribution (manuscript in preparation). In particular, we use the CloudSat cloud classification [SW08] in conjunction with our own classification of CALIPSO clouds to characterize the interannual variability of CFO as a function of cloud type. From a subset of these cloud profiles we employ a method similar to [YFH10] to determine which clouds have the greatest impact on the radiative balance of the tropical atmosphere.
CHAPTER 2

Developing an Integrated Water Vapor Data Set

“Two are better than one, because they have a good return for their labor...” – King Solomon

2.1 Preamble

Historically, quantification of the UTLS H$_2$O budget has been hindered by the lack of global high-resolution measurements of H$_2$O throughout the troposphere and stratosphere. However, recent studies have quantified the global upper tropospheric H$_2$O structure with the global soundings provided by the “Afternoon Constellation” (A-Train) [SVB02] Atmospheric Infrared Sounder (AIRS) [ACG03] flying on NASA’s Aqua satellite platform [GFE06, GWF04, KLE08, KGF09]. On the same constellation the Microwave Limb Sounder (MLS) [WFH06], located on Aura, provides global UTLS H$_2$O soundings [RLB07].

The instruments together provide high quality UTLS H$_2$O measurements, however, AIRS primarily has sensitivity to H$_2$O consistent with concentrations from the boundary layer up to the UT. The lower threshold of AIRS sensitivity to H$_2$O was derived empirically, in [GWF04], to be $\sim$10 ppmv based on comparisons with aircraft measurements. [FRW08] derived a H$_2$O minimum threshold to be between 20–30 ppmv. On the other hand, MLS has sensitivity in the stratosphere down to the UT with sensitivity to values between $\sim$0.1–1 ppmv; MLS does have
an saturation threshold of $\sim 500$ ppmv [RLB07]. Both [RLB07] and [FRW08] determined that AIRS and MLS agreed best between $\sim 250-260$ hPa.

For this work, we quantify the relative sensitivity of both instruments to UTLS H$_2$O directly from each instruments averaging kernels. A prior study by [MB08] computed AIRS AK’s for selected sites and approximated the instrument resolution for temperature and H$_2$O. We use the AK’s for an entire year of AIRS retrievals to evaluate, on a global scale, what AIRS’ sensitivity is to UTLS H$_2$O. This information provides the means to merge AIRS and MLS H$_2$O retrievals into a single profile which spans the troposphere and stratosphere.

2.2 Data and Methodology

2.2.1 Atmospheric Soundings

The AIRS version 5 support product [SBB06] and MLS version 2.2 retrievals [LSR06] are co-located for 2008 in the 40 S–40 N latitude band according to [FRW08]. Co-location is simply defined as the nearest AIRS footprint to a given MLS footprint (the approximate lag time between Aura to Aqua measurements is about 15 min, but due the MLS viewing geometry the lag is reduced to 8 min). The nearest neighbor co-location is implemented in order to preserve the AIRS resolution. The AIRS reports H$_2$O mixing ratios as a mean layered quantity between adjacent pressure levels while MLS reports retrievals on a 12 levels per decade change in $\log_{10} P$. Since the MLS reports H$_2$O as a pressure level quantity, we redefine the AIRS H$_2$O concentrations as a geometric mean quantity between adjacent pressure levels. These values are then interpolated ($\log(P)$ vs. $\log$(H$_2$O)) to a hybrid pressure grid that is a combination of the AIRS and MLS grids. The difference in horizontal resolution are not accounted for; the vertical
resolution of both instruments are similar (∼2–3 km).

Figure 2.1a is an example of a single AIRS and MLS pair of co-registered H\textsubscript{2}O profiles at 0.17 N and 176 W. When comparing large sets of profiles, one persistent characteristic is that they generally do not resemble each other at pressures higher than ∼300 hPa; this is well documented in [RLB07] and [FRW08]; this is the case for many other profiles we have analyzed as well. At pressures 100 ≤ P ≤ 260 hPa, in the UT (Fig. 2.1b), the profiles become more similar. However, as seen in Fig. 2.1b, they do differ enough that a discontinuity exists between the profiles.

One goal of this study is to merge them together in a consistent and rigorous manner in order to form H\textsubscript{2}O profiles that span the entire atmospheric column. The “Joined” curve in Fig. 2.1b is a sample merged profile combining the AIRS (red) and MLS (blue) curves. The merging process is discussed in detail in Section 2.3.2 and requires a quantitative estimate of AIRS and MLS sensitivity to H\textsubscript{2}O in the UTLS. This is accomplished with the use of instrument- and retrieval algorithm-specific AK’s.

### 2.2.2 Optimal Estimation Retrievals

Satellite retrievals of an atmospheric constituent such as H\textsubscript{2}O are typically done using an optimal estimation framework. The retrievals can be linearly related to the actual H\textsubscript{2}O distribution in the following manner:

\[
\hat{x}_i = x_{0,i} + A_i(x - x_{0,i}) + G_i n_i
\]  

(2.1)

where \(A_i\), \(x\), \(\hat{x}_i\), \(x_{0,i}\), \(G_i\), and \(n_i\) are the averaging kernel, truth profile, retrieved profile, a-priori profile, gain matrix, and noise vector for the \(i\)th instrument. \(A_i\) is an \(N \times N\) matrix where \(N\) is the number of retrieval levels. The product \(G_i n_i\) represents the errors introduced by the AIRS/MLS measurement and forward
Figure 2.1: (a) Sample co-located tropical H$_2$O profiles for AIRS (red) and MLS (blue) at 0.17N and 176W, (b) the H$_2$O profile zoomed in at the levels corresponding to the peaks in the averaging kernels in (c) for AIRS (red), MLS (blue), and a merged version of the profile (see text for details). Panel (c) shows a sample set of tropical averaging kernels from MLS (solid colored lines) for the pressure levels 83, 100, 121, 147, 178, 215, 261, and 316 hPa; dash-dotted black line is the verticality (values at top abscissa). Panel (d) same as (c) but for AIRS (at 0.17N and 176W). Panel (e), same as (d), but at 82S and 90W.
model. Now, $G_i$ can be written as

$$G_i = (K_i^T S_{e,i} K_i + S_{a,i})^{-1} K_i^T S_{e,i}^{-1}$$

(2.2)

$$A_i = G_i K_i$$

(2.3)

where $K_i$, $S_{e,i}$, and $S_{a,i}$ are the Jacobian, error covariance, and a-priori covariance matrices for the $i$th instrument. Equations 2.1 and 2.2 are the basis for optimal estimation retrievals [Rod00]. Note the gain matrix in Eq. 2.2 takes into account the influence of the a-priori covariance on the final retrieved profile and also impacts the AK matrix (Eq. 2.3).

In order to compare the AK’s of AIRS and MLS (or between any two remote sensing systems) the influence of the a-priori should be accounted for in order to compare the AK’s in a consistent manner. Unfortunately, the current AIRS retrieval does not work in an optimal estimation framework and does not incorporate any effects from the a-prior covariance matrix $S_a$, thus making it virtually impossible to account for any AK differences between the AIRS and MLS a-prioris. Although both the AIRS and MLS retrievals do use Eq. 2.2 in some form, the AIRS retrieval, however, uses a matrix $H$ in place of $S_a$ to prevent the matrix inversion from failing. Since the gain matrices, which account for instrumental and retrieval errors, do not have the same information, i.e. only MLS accounts for the a-priori covariance, the interpretation of comparisons between the AIRS and MLS AK’s needs to be handled with care. Nevertheless, as will be shown in Section 2.3.1 and 2.3.2, the AIRS and MLS AK’s do provide consistent information that quantifies their relative sensitivity to UTLS H$_2$O.
2.2.3 AIRS averaging kernels

The AIRS AK’s are derived on a profile by profile basis at a horizontal spatial resolution of ∼45 km; they are computed using the methodology described in [MB08]. The MLS consists of a single set of AK’s for this latitude band. This is because the limb geometry of MLS senses through a large amount of atmosphere, for a given layer, giving it a high sensitivity to H\textsubscript{2}O, though with a reduced spatial resolution of about ∼200 km along-track. In the case of MLS, the sensitivity is primarily driven by the viewing geometry, so that the AK’s do not change much from footprint to footprint. Therefore, herein we focus our analysis on the characterization of the AIRS AK’s.

One application of AK’s is to degrade high vertical resolution correlative measurements, such as radiosondes of temperature and H\textsubscript{2}O for comparison with the more coarsely resolved satellite data [MB08]. The following equations provide the formalism to do so:

\begin{align*}
\hat{x} &= \hat{x}_0 + \Phi(x - \hat{x}_0), \\
\Phi &= \frac{\partial \hat{x}}{\partial x} \tag{2.5}
\end{align*}

where the $\hat{x}$, $\hat{x}_0$, $\Phi$, and $x$ are the smoothed version of the true atmospheric profile, a priori, the AK matrix, and the true atmospheric profile, respectively [Rod00].

Again, in an optimal estimation framework, $\Phi$ would be a $N \times N$ matrix where $N$ represents the number of retrieval levels. However, the AIRS retrieval uses a reduced measurement space of 12 free parameters to retrieve H\textsubscript{2}O for 100 levels [MB08]. This results in an $11 \times 11$ $\Phi$ matrix. In order to construct an AK matrix with $N \times N$ elements ($N = 100$ levels), we construct an “effective” AK using the methodology described in [MB08]. Briefly, the retrieval basis functions
in the AIRS retrieval system are 11 trapezoidal functions defined over 100 retrieval levels, giving a $100 \times 11$ matrix $F$. The effective AK matrix computation is shown in Eq. 2.6:

$$A = F \Phi F^+$$

(2.6)

where $F^+$ represents the adjoint of $F$. $A$ is now a $100 \times 100$ matrix that now describes the AK behavior on the AIRS retrieval levels rather than the reduced measurement space [MB08].

The AK matrix provides three pieces of information [MB08]: (1) the vertical distribution of information content received from the instrument’s radiances, (2) the verticality, and (3) an approximation of the instrument’s vertical resolution; the full-width half maximum (FWHM) of the AK peak is one way to estimate resolution. The vertical distribution of information tells us where the retrieval information content is coming from as a function of pressure. The verticality, which is the sum of the $i$th row in $A$, tells us how much total information is gathered from the radiances. A verticality near unity indicates the retrieval, for a particular level, comes primarily from the measured radiances, while low verticalities indicate more influence from the a priori.

Figure 2.1c shows the set of MLS AK’s. Notice that all the kernels peak at the correct pressure levels. Note also the verticality (dash-dotted line) is nearly unity at all pressure levels as well. Figure 2.1d and e show a set of tropical and high latitude AK’s from AIRS. The lowest pressure level (highest in altitude) AIRS peaks at (Fig. 2.1d) is around 260 hPa (there is a peak around 170 hPa, but the verticality for this level is $\sim 0.65$, much lower than for higher pressure levels). The MLS verticality indicate that its information content primarily comes from the measured radiances at all levels while AIRS verticality decreases sharply beyond 260 hPa. Although the AIRS verticalities drop off at pressures below 260 hPa, the
verticalities still indicate AIRS has some sensitivity to H\textsubscript{2}O depending on mixing ratio and local temperature lapse rates; this will be discussed in more detail in Section 2.3.1. A sample extra-tropical sounding shows verticality that peaks around 400 hPa then rapidly drops off at lower pressures (Fig. 2.1e). As will be shown, this sharp drop in information content occurs for all AIRS H\textsubscript{2}O retrievals throughout the tropics and extra-tropics (at higher latitudes the decreases begin further down in the troposphere; see Fig. 2.1e).

2.3 Results

2.3.1 Averaging kernel statistics

One year (2008) of co-located AIRS and MLS profiles is analyzed in a 40 S–40 N latitude band, limited to AIRS profiles with a quality flag of “PGood=Psurf” where “Psurf” is the surface pressure (see [te10a] for quality flag details). Figure 2.2 summarizes the statistical findings. Each row represents a particular retrieval pressure level. In this study pressures between 83 hPa (lower stratosphere) and 407 hPa (free troposphere) are analyzed. Column 1 shows the vertical distribution where the AK peaks for the aforementioned retrieval levels; column 2 shows the distribution of verticality, herein termed the \textit{total verticality} (TV); and column 3 shows the verticality but for a narrow set of layers encompassing the retrieval level of interest; herein termed the \textit{local verticality} (LV). This narrow layer spans six atmospheric levels on either side of the retrieval location, totaling 13 levels (out of the 100 possible levels) that go into computing LV (remember that TV is computed summing the AK values for all 100 pressure levels).

From column 1 of Fig. 2.2 it can be observed that AIRS AK’s generally peak at or near the retrieval level up to the 260 hPa level. This suggests that most of
the retrieval information content comes from the correct parts of the atmosphere up to 260 hPa. Moving deeper into the UT and transitioning to the tropical tropopause layer (TTL) region [FDD09] (70-150 hPa), the kernels generally peak at 260 hPa and at 170 hPa. In fact, from 170 hPa to 83 hPa, the kernel peak distribution does not change much at all, indicating that AIRS has little to no sensitivity to H$_2$O in this region of the atmosphere. Note that the AK’s never peak near the 212 hPa level indicating that AIRS has a “blind spot”, i.e. weaker sensitivity in that location of the atmosphere.

Transitioning to column 2 of Fig. 2.2 one can see that in addition to the AIRS retrievals sensing the expected region in the free troposphere to the lower parts of the UT (260–407 hPa), the retrievals also receive most of the information content from the measured radiances, i.e. TV near unity. From about 212 hPa and upwards the TV drops off rapidly suggesting that AIRS has reduced sensitivity to H$_2$O in the UTLS. One thing to note is that even though TV may be near unity, this does not imply strong sensitivity to H$_2$O for a particular level. The TV provides a cumulative representation of the information content of the retrieval. It does not indicate which vertical layer(s) the information comes from. Column 3 serves to draw out this missing information.

The LV, shown in column 3 of Fig. 2.2, represents the verticality for a relatively narrow layer around the retrieval level. This parameter serves to extract the fraction of the TV that comes from this narrow band of pressures. AIRS can be considered highly sensitive to H$_2$O for a particular level when LV is a significant fraction of TV as long as TV (quantified in column 2) is near unity. However, if LV is a small quantity compared to its corresponding TV, this suggest a high degree of correlation between multiple layers far removed from the retrieval level. Note, LV is computed for MLS and the values are a significant fraction of the
MLS TV \( (\frac{LV}{TV} \geq 0.9) \).

The mean relative difference of all LV to all TV, \( \frac{1}{N} \sum \frac{LV}{TV} \) (for \( N \) measurements, reported as a percent), is shown as the first value in each parenthesis (column 3). These percent differences quantify the fraction of the total verticality within in a narrow layer around the retrieval level of interest. LV between 407–314 hPa falls off by only \( \sim 13\% \). The 260 hPa level shows first indication of LV falling off abruptly (30%). This, is due in part to the TV distribution being slightly skewed to higher values than at higher pressures. Also, the mean TV at 260 hPa is \( \sim 10\% \) greater than the mean TV for higher pressure levels; the mean LV at 260 hPa is \( \sim 0.8 \). Above this level the mean LV drops off rapidly, and the distribution of LV shifts quickly to lower values as well, suggesting a loss of sensitivity to H\(_2\)O in the TTL region and above.

The second set of values in the parenthesis represent the mean percent change in LV relative to the mean LV at 407 hPa. The average LV from 407 hPa to 260 hPa only falls off by \( \sim 9\% \), suggesting that AIRS is sensitive to H\(_2\)O at these levels. Note that at 314 hPa the LV distribution is broader than its TV counterpart, indicating AIRS is losing some ability to discriminate between levels around this layer and the a priori having some influence on the retrieved profile. Nevertheless, the information content is still very high, giving us confidence in the ability of AIRS to measure H\(_2\)O in the lower parts of the UT in the tropical and extra-tropical atmosphere. The region between 83–212 hPa, which encompasses the TTL, is marked by a mean LV drop from \( \sim 39\% \) to \( \sim 98\% \), clearly showing that the information content at these retrieval level drops off rapidly. The higher altitude TV distributions suggest that there is some information gleaned from the measured radiances, i.e. non-zero TV, however, this information is mainly obtained from the 170 hPa and 260 hPa levels (see column 1 of Fig. 2.2) essen-
tially indicating that the AIRS algorithm is spuriously moving information from lower levels to higher levels where it has little to no sensitivity.

One can argue that AIRS provides high quality H₂O retrievals in the tropics and extra-tropics for pressures as low as ~260 hPa, or perhaps to lower pressures in moister conditions with discernible vertical thermal gradients. This, however, is not true for higher latitudes. Figure 2.3 shows the same statistics as shown in Fig. 2.2 for all latitudes outside the 40S to 40N latitude band. Figure 2.3 reveals that AIRS loses sensitivity starting at around the 314 hPa level. This is seen by the abrupt drop in the mean LV by about 22% relative to the LV at 407 hPa, and by the broadening of the TV and LV distributions starting at the 314 hPa level. This broadening, that also occurs at lower pressures, indicates that retrievals at these level are strongly influenced by the a priori. The UTLS kernel peak distributions also peak at many more levels outside this region of the atmosphere (column 1).

In general, AIRS is sensitive to regions with relatively higher concentrations of H₂O (e.g. the tropical atmosphere) and where there is a thermal lapse rate. However, it is important to determine if the averaging kernels reflect this assumption. Figure 2.4 shows probability distribution functions (PDF’s) of H₂O (columns 1 and 2), temperature lapse rate (dT/dP, column 3) and LV (column 4) at each retrieval level partitioned by various levels at which the AK’s peak. Column 1 shows PDF’s for the final AIRS retrievals while column 2 shows H₂O distributions of the a priori passed into the AIRS retrieval algorithm (in actuality the a priori in the version 5 dataset are the retrieval profiles after one iteration of the retrieval algorithm). Each curve (in all columns) represents a PDF of values that correspond to where the kernels primarily peak for each retrieval level; each graph shows a set of three retrieval levels that correspond to the most statisti-
Figure 2.2: Distributions of: column 1) locations of where the averaging kernels peak, column 2) total verticality, and 3) local verticality for each retrieval level at pressures 83, 103, 126, 142, 170, 212, 260, 314, 375, and 407 hPa. Data are for AIRS data only. The total number of profiles is 384,017, so all distributions in column 1 and subsequent columns are normalized to this quantity.
Figure 2.3: Same as Fig. 2.2 but for latitudes polewards of 40°S to 40°N. Total number of profiles is 360,820.
cally significant pressure levels shown in column 1 of Fig. 2.2. For each retrieval level, the red curves show distributions that peak near or at the retrieval level of interest (at 170 hPa the curve is black). We focus on the region where AIRS has weak to strong sensitivity (Fig. 2.2) to H₂O (170 hPa to 407 hPa).

Comparing column 1 and column 2 in Fig. 2.4 one finds that the H₂O PDF’s noticeably differ. This reflects the high LV (column 4) and TV (column 2 of Fig. 2.2) values. At most of these levels (pressures greater than 212 hPa) the LV reflects that the majority of the data is not largely governed by the a priori information content. Looking more closely at column 1 the curves corresponding to the retrieval level, or nearest the retrieval level, are more moist than than distributions with peak kernel values differing from the retrieval height (when the AK’s peak at the correct pressure level the curves are plotted in solid bold lines). This indicates that AIRS has higher sensitivity to H₂O in more moist atmospheric conditions. The blue and black curves represent distributions of H₂O corresponding to situations when the AK’s peak at some layer below or above the retrieval level, respectively. For 170 hPa the red and blue curves are both at higher pressures than the retrieval level because the AIRS AK’s never peak at pressures lower than 170 hPa. The LV distributions (column 4 of Fig. 2.4) follow the H₂O distributions, i.e. moister distributions correspond to higher LV values.

Column 3 shows that the temperature lapse rate distributions closest to or at the retrieval levels, in general, have larger gradients than the distributions at levels at other pressure levels; this is particularly true between 170–314 hPa. This is not surprising as the H₂O IR signature is not only dependent on the concentration of H₂O, but also on the presence of a vertical thermal gradient.
Figure 2.4: Distributions of: column 1) H$_2$O for the final retrieval, column 2) H$_2$O for the initial guess (a priori), column 3) temperature lapse rate as a function of pressure (note this is $-dT/dP$), and column 4) local verticality for each retrieval level partitioned by the pressure levels where the kernels peak for each retrieval level. Total number of profiles is the same as in Fig. 2.2. Colors indicate various pressure levels (in hPa) shown in column 2.
2.3.2 Merging the Profiles

The AIRS AK’s were shown to have skill in identifying the pressure levels where it can and cannot retrieve H$_2$O well. The fact that previous inter-comparisons confirmed that the AIRS and MLS agree best between 250–260 hPa [RLB07, FRW08], also indicates the AIRS AK’s do accurately quantify its sensitivity to H$_2$O. Since AIRS and MLS both have sensitivity to H$_2$O and agree best around 260 hPa it is conceivable to merge their profiles with 260 hPa acting as an anchor point between the AIRS and MLS profiles. This requires a method to smoothly transition between the AIRS and MLS profiles and is described below.

We briefly note that in an optimal estimation framework the AIRS and MLS AK’s are influenced by their respective covariances and a-priori covariances. These influences, in principle, need to be accounted for before one can even use the AK’s from both instruments to quantify their sensitivities in a consistent manner. Thus, the a-priori differences need to be considered in order to merge the data between the two systems. However, because AIRS does not work in an optimal estimation framework, quantifying any differences that result from their a-priori and covariances is virtually impossible. However, the results of Section 2.3.1 indicate that the AK’s of both instruments do capture their sensitivities reasonably well. Below, we describe a simple method, using the AIRS and MLS AK’s, that is able to produce merged profiles which, within their error estimates, do not deviate from either instrument’s interpretation of the atmospheric state.

The AIRS and MLS data products are provided on different pressure grids, therefore one needs to compute the AIRS AK on the MLS levels in order to produce comparable AK. To merge the profiles the AIRS AK’s are computed on the MLS levels using Eq. 2.6. This is necessary as the 12 levels per decade for MLS leads to a coarser pressure grid than the AIRS L2 support product levels in
the UTLS. In this case, instead of defining the trapezoidal basis functions on the 100 retrieval levels of AIRS, we combined the MLS pressure levels with the AIRS L2 grid to compute the basis function matrix $F$ in Eq. 2.6 with $[100 \times (\text{AIRS}) + 47 \times (\text{MLS})] \times 11$ elements. Then the basis functions only on the MLS pressure levels (producing a new $F$ with dimension $47 \times 11$) are extracted and inserted into Eq. 2.6 to compute a $47 \times 47$ “effective” AK matrix (remember $\Phi$ is an $11 \times 11$ matrix). This procedure, in effect, redistributes the information content in the $100 \times 100$ AIRS AK’s matrix onto a $47 \times 47$ matrix corresponding to the 47 MLS pressure levels, resulting in two AK matrices, $A$ and $M$, that represent the AIRS and MLS AK’s respectively. A successful remapping of the AK matrix to the MLS levels satisfies the following condition: $Tr(\Phi) = Tr(A)$ where $Tr$ is the trace of the matrices. The AIRS $H_2O$ are also interpolated ($\log(P)$ vs. $\log(H_2O)$) to the MLS levels.

After computing $A$ and $M$ the local verticalities (LV) are computed from these AK matrices using AK values from the $i$, $i-1$, and $i+1$ pressure levels for AIRS and MLS, approximating the FWHM of the MLS AK’s. The AIRS $H_2O$ profiles are kept from 1013 hPa up to 261 hPa (based on the statistics in Fig. 2.1), and LV is computed from 215 hPa up through the rest of the atmospheric column. All LV serve as weighting coefficients for merging the AIRS and MLS $H_2O$ profiles. The final hybrid pressure grid comprises of MLS pressure levels from the top of the atmosphere down to 215 hPa and AIRS pressure values from 261 hPa down to 1013 hPa.

To finally merge the AIRS and MLS profiles a weighted mean between the AIRS and MLS $H_2O$ values is computed, starting from 215 hPa and upwards in altitude, using the computed LV or weighting coefficients in the following manner:

$$q_{i}^{\text{join}} = \frac{q_{i}^{A} LV_{i}^{A} + q_{i}^{M} LV_{i}^{M}}{LV_{i}^{A} + LV_{i}^{M}},$$

(2.7)
where \( q_i^A \), \( q_i^M \), and \( q_i^{\text{join}} \) are the \( \text{H}_2\text{O} \) concentrations at the \( i \)th retrieval level for AIRS, MLS, and the newly merged profile, respectively. \( LV_i^A \) and \( LV_i^M \) denote LV’s for AIRS and MLS, respectively. The black dotted line in Fig. 2.1b is a sample resultant profile from merging the MLS (blue) and AIRS (red) profiles. One observation is that the merged profile has no discontinuities anywhere near the merge region. Furthermore, as will be the case for all merged profiles, it is constrained by the AIRS and MLS \( \text{H}_2\text{O} \) values, i.e.:

\[
q_i^A \leq q_i^{\text{join}} \leq q_i^M \quad \text{or} \quad q_i^M \leq q_i^{\text{join}} \leq q_i^A.
\]

Figure 2.1b demonstrates that the selected co-located profiles can be joined in this manner to create a smooth continuous \( \text{H}_2\text{O} \) profile.

In order to quantify if this technique, at a minimum, produces smooth functions for all merged profiles, the parameter depicted in Fig. 2.5a–d is computed. The parameter,

\[
DQ = \frac{1}{\delta P} \frac{\delta q}{q_{i+1}}
\]

(2.8)

with \( \delta q = q_i - q_{i+1} \), represents the percentage change in \( \text{H}_2\text{O} \) between the \( i \) and \( i+1 \) level, scaled by the pressure change \( \delta P \) (\( \delta P = P_i - P_{i+1} \)) to remove the effect of having a varying pressure grid. This can also be interpreted as a distribution of \( \text{H}_2\text{O} \) lapse rates with respect to pressure, reported as a fractional deviation from the \( i+1 \) retrieval level. \( DQ \) is used in order to more clearly identify possible unphysical “kinks” in the profiles. The colors in the distribution (Fig. 2.5a–c) denote the frequency of occurrence of lapse rates for the entire 2008 data record.

The quantity \( DQ \) is shown for the original MLS (Fig. 2.5a) and AIRS (Fig. 2.5b) profiles. One observation is the AIRS \( DQ \) is smooth throughout the range of the UTLS (the region where AIRS and MLS overlap in sensitivity), while MLS shows
a wider variety of lapse rates. Figure 2.5c shows the PDF’s of lapse rates for the merged data. The distributions from the higher pressures become slightly narrower due to the influence of the AIRS H$_2$O values. However, as we move to lower pressures, the distributions start to mimic the MLS values because LV$_i^A$ is small at these levels; Fig. 2.5d depicts this more clearly.

At 261 hPa the JOIN (green) and AIRS (red) data are identical because we selected this level as the lowest pressure level that gives full weight to AIRS. The JOIN and MLS curves (blue) already start to converge at 215 hPa. The AIRS and JOIN values at 178 hPa are more similar for a couple reasons: 1) LV$_i^A$~0.5–0.6 at these levels giving AIRS noticeable influence on the mean profile especially for the moist profiles that have greater values of LV, and 2) the AIRS AK’s often peak around 170 hPa (Fig. 2.2 column 1), while never peaking at 212 hPa. This also explains the similarity of the JOIN and MLS distributions at 215 hPa. Although the distribution of DQ$_{178}$ for AIRS might resemble that of the joined profiles, one needs to remember DQ represents a relative percent change in the H$_2$O lapse rate which only measures the smoothness of the profiles as a whole, providing no qualitative measure of how well mixing ratios compare at any given level [RLB07, FRW08].

From 121 to 83 hPa, the DQ distributions for MLS and the merged data look nearly identical. By the time we reach 83 hPa, the JOIN and MLS curves are virtually identical and have no resemblance to the AIRS DQ distribution. From Fig. 2.2 we see that the LV$_i^A$ distribution at 83 hPa is strongly skewed towards zero while LV$_i^M$ (see Fig. 2.1c) is near unity at these levels. Note the total number of high quality merged profiles is limited by the number of usable MLS profiles, leading to fewer merged profiles than the number of high quality AIRS profiles noted in Fig. 2.2.
Figure 2.5: Joint distribution for the value $DQ=\frac{1}{\delta P} \frac{\delta q}{q_{i+1}}$ (multiplied by 100 to get %) all (a) MLS, (b) AIRS, and (c) joined $H_2O$ profiles. Color scale shows frequency in %. Panel (d) shows the individual PDF's for the joint distributions in (c) for pressures of 83, 100, 121, 147, 178, 215, and 261 hPa. Frequency units are in % of the total number of profiles shown in the bottom panel in (d). Total number of profiles is 367,689.
Even though the profiles are smooth, we also need to determine if the merged profiles preserve what AIRS and/or MLS interprets as the atmospheric state within the observational uncertainty of AIRS and MLS. Since the merged profiles are constrained by the AIRS and MLS H$_2$O concentrations and averaging kernels, the merged profiles at a “zeroth” order level cannot deviate far from either instrument’s retrievals. However, characterizing the robustness of this dataset requires determining the merged profiles uncertainty in relation to the stated uncertainties of AIRS and MLS. For pressure levels $P \leq 121$ hPa we expect the MLS uncertainties to be the appropriate metric since the AIRS averaging kernels at these upper levels do not contribute much to the merged profile H$_2$O concentrations since $LV^A_i < 0.25$ and $LV^M_i \sim 1$. However, for larger pressures down to 215 hPa we expect that both the AIRS and MLS uncertainties will be important constraints on the performance of the merged profiles since $LV^A_i$ and $LV^M_i$ are both non-negligible.

There are only a few studies that quantify the accuracy and precision of AIRS. [DBG06] estimated the AIRS accuracy and RMS to be 25% and 40% at 150 hPa and 200 hPa from comparisons with radiosondes launched from 538 stations around the earth using a limited number of stations between the 40 S–40 N band. [TRK06] estimated the AIRS RMS error (between 100–266 hPa) from comparisons with radiosonde measurements over the tropical western Pacific to be between 20–25%. Since these values are still not well constrained, because of the limited spatial coverage of the inter-comparisons, a fixed value of 25% for both the accuracy and precision is used for all levels in our computation. We do note that these estimates are likely high as the averaging kernels were not available to smooth the radiosonde data to the AIRS resolution. The MLS stated accuracies of 7%, 8%, 12%, 15%, 17%, and 25% for 83 hPa, 100 hPa, 121 hPa, 147 hPa, 178 hPa, and 215 hPa; and the stated precisions of 15%, 10%, 15%, 20%, 20%,
and 40% for the same levels are used [RLB07].

Figure 2.6 shows the PDF’s (in %) of \((q^{\text{join}} - q^{\text{AIRS,MLS}})/q^{\text{join}}\) for the aforementioned levels. Shown in each panel are the root sum square error (RSSE), i.e. \(RSSE = \sqrt{\sigma_a^2 + \sigma_p^2}\) (\(\sigma_a\) and \(\sigma_p\) are the accuracy and precision, respectively), for MLS (solid black vertical lines) and AIRS (dashed-dot vertical lines). The percentage of \(q^{\text{join}}\) that fall within the MLS uncertainties (RSSE=17%, 13%, 19%, 25%, 26%, and 40%) or AIRS uncertainties (RSS=35% for all levels) are 99.9%, 99.8%, 94.9%, 84.7%, 96.6%, and 88.7%. For 83 hPa and 100 hPa we expect that most of \(q^{\text{join}}\) will fall within the uncertainties since H\(_2\)O concentrations at these levels are almost completely influenced by MLS. For the levels below, we still have \(q^{\text{join}}\) that fall within the either instruments uncertainty estimates for at least \(\sim 85\%\) of the time. Therefore, we can conclude that the newly constructed H\(_2\)O profiles do not deviate significantly (more than 1 standard deviation) from the original retrievals within their stated uncertainties. For the values that do fall outside the error estimates, it is noted that they are constrained by either instruments H\(_2\)O concentrations, weighted by LV. Thus, even these values do not deviate far from either instruments retrievals.

In order to robustly validate the merged profiles one would need to compare these profiles with a large set of radiosonde measurements that capture a variety of atmospheric conditions. Subsequently, one would need to also apply AK’s of both the AIRS and MLS instruments to account for the coarser remote sensing resolution. It is unclear how to apply the AK’s for this application because the new profiles represent a weighted mean of AIRS and MLS for \(P \leq 215\) hPa. However, even without a detailed inter-comparison with in-situ measurements, the merged profiles do not deviate far from the AIRS and MLS soundings within their stated uncertainties for a majority of the profiles.
Figure 2.6: Cumulative frequency (abscissa) distribution of the % difference \( \frac{q^{\text{join}} - q^{\text{MLS}}}{q^{\text{join}}} \) (black curve) and \( \frac{q^{\text{join}} - q^{\text{AIRS}}}{q^{\text{join}}} \) (grey curve) for 83, 100, 121, 147, 178, and 215 hPa (abscissa). Solid black and dashed-dotted grey vertical lines are the root-sum square (RSS) of the accuracy and single profile precision estimates for MLS and AIRS, respectively. Ordinate is normalized frequency.
Figure 2.7: Mean maps (July 2008–September 2008) of H$_2$O concentrations binned in $4^\circ \times 4^\circ$ grid cells globally between $40^\circ$S–$40^\circ$N for: (a) 147, (b) 215, (c) 261, and (d) 300 hPa. Color scales are in units of ppmv.
2.3.3 Mean H$_2$O Maps

In Section 2.3.2 a combined AIRS/MLS merged profile was constructed. It was shown that the profiles are smooth and do not substantially deviate from the AIRS and MLS soundings within their instrumental uncertainties. Figure 2.7 shows global maps (July–September, 2008) of H$_2$O concentrations for 147, 215, 261, and 300 hPa between 40$^\circ$S–40$^\circ$N. All maps feature a moist UT due to the convection associated with the Indian monsoon. Also apparent are local maxima over land due to convective uplift. These maps compare well with the maps [FRW08] (their Fig. 3), and further support that our new merged dataset does not change the instrumental interpretation of the atmospheric state.

2.4 Discussion

We have quantified the sensitivity of the Atmospheric Infrared Sounder (AIRS) to H$_2$O in the upper troposphere/lower stratosphere (UTLS). For soundings between 40$^\circ$S–40$^\circ$N AIRS can reliably quantify H$_2$O up to $\sim$260 hPa. At lower pressures AIRS sensitivity rapidly drops in relatively drier conditions and somewhat slowly in more moist conditions. At higher latitudes AIRS starts to lose sensitivity at even higher pressures (around 375 hPa). This is a consequence of increased sensitivity in atmospheres with high concentrations of H$_2$O and larger vertical thermal gradients. The fact that AIRS has high information content (i.e. high total verticality (TV) and local verticality (LV)) up to 260 hPa corroborates previous results that show AIRS and the Microwave Limb Sounder (MLS) best agree between 250–260 hPa. Thus, this indicates that the AIRS AK’s have skill in quantifying its sensitivity to H$_2$O.

The merged profiles in this study correspond to the nearest AIRS pixel to each
MLS point. The method described in this study has also been used to merge the profiles by co-locating the three nearest AIRS pixels (at \( \sim 45 \) km resolution) to each MLS point with \( \sim 200 \) km resolution. This produces a dataset that preserves the native AIRS resolution from the surface up to 261 hPa where \( \text{H}_2\text{O} \) gradients are much stronger than in the TTL. Statistics similar to those shown in Fig. 2.6 were computed on two pixels adjacent to the nearest neighbor point and yielded very similar results.

Since the AIRS and MLS averaging kernels (AK’s) have quantitative skill, the information gleaned from the AIRS and MLS AK’s helped to produce the first merged \( \text{H}_2\text{O} \) measurement record that spans the entire troposphere and stratosphere. We have also quantified that the merged profiles do not significantly deviate from the original AIRS and MLS soundings and largely remain within their respective uncertainties. It should be noted that the method used to construct these profiles cannot be applied for all trace gas retrievals. This procedure is valid for \( \text{H}_2\text{O} \) because its concentration falls off exponentially (e-folding height of \( \sim 2 \) km) and does not have strong gradients and sign reversals in the UTLS as compared to other gases such as ozone, which sharply increase in the lower stratosphere.

Lastly, we do not suggest this methodology is the optimal method to merge these datasets. A complete treatment of the problem would utilize a joint, simultaneous optimal estimation retrieval with the AIRS and MLS radiances (e.g. jointly solving Eq. 2.1 for both AIRS and MLS). This problem requires much more research on the practical considerations of a joint instrument retrieval algorithm that accounts for the vastly different instrument sensitivity and sampling characteristics between AIRS and MLS. What we have done is develop a method that takes advantage of the overlapping vertical sensitivity of AIRS and MLS and
the typical smoothly varying nature of H$_2$O in the UTLS. The newly constructed profiles now offer the unprecedented opportunity to explore processes that interact between the troposphere and stratosphere in a self-consistent dataset.
CHAPTER 3

Tropical Interannual Variability of Temperature and Water Vapor

“The moments of happiness we enjoy take us by surprise. It is not that we seize them, but that they seize us.” – Ashley Montagu

3.1 Preamble

The processes that govern the tropical upper-tropospheric (UT) water vapor (H$_2$O) structure determines the final H$_2$O concentration that enters into the tropical lower-stratosphere (LS). The tropical UT radiative structure is most sensitive to changes in H$_2$O relative to any other vertical location in the tropics [HS00, SHC07]. However, the stratospheric H$_2$O distribution has also been shown to have a non-negligible impact on climate [SRP10]. Many different tropical processes contribute to the H$_2$O distribution including convective moistening [Sod04, HS07], dehydration [FBH05, FH05], horizontal advection through cold trap regions, precipitating thin cirrus [HG01], and convective mixing [SD01, SD03]. However, the aforementioned processes also impact convection on interannual timescales (~1-5 years). The primary tropical modes of variability with consistent timescales are the El-Niño Southern Oscillation (ENSO) [Phi90, WRM98] and the Quasi-
Biennial Oscillation (QBO) [Ree65b, Ree65a, Ham84, Dun85]. [GRM01] showed that the ENSO modulates tropopause temperatures in the TCP, making them colder (warmer) during El Niño (La Niña) periods.

The QBO manifests itself as descending temperature anomalies [AHL87] in thermal wind balance with the descending westerly/easterly wind anomalies, which are a result of the interaction between the zonal mean wind and the upward propagating equatorial waves (from tropical convection) in the stratosphere [BGD01]. [ZGZ01b] showed that when these descending temperature anomalies were in phase with the La Niña (El Niño) the dehydration potential of the tropical tropopause layer (TTL) increased (decreased). Work by [FBH05] and [FH05] showed that TTL H\textsubscript{2}O concentrations can be reconstructed, though with some uncertainty, tracing a parcels history back to the coldest temperature it experiences–this is termed the Lagrangian Cold Point (LCP)–and that the interannual modes of the ENSO and QBO both impact the LCP.

Although the previous studies showed the possibility of the joint impact of the ENSO and QBO on the temperature and H\textsubscript{2}O distribution, they did not explicitly observationally show the measured impact of the ENSO and QBO on the H\textsubscript{2}O distribution. With this as our motivation, we construct a ~6.5 year temperature and H\textsubscript{2}O record of co-located measurements by AIRS and MLS. From this record we characterize the annual cycle and the interannual cycle of these quantities. The interannual variability is further quantified by characterizing the different impacts of the ENSO and QBO depending on the relative phases of these modes of variability.
3.2 Data and Methodology

3.2.1 A-Train Soundings

The H$_2$O profiles are taken from two sources: 1) the AIRS [ACG03] and 2) the MLS [WFH06] which fly on the Aqua and Aura spacecrafts respectively. The AIRS, which flies on a sun-synchronous orbit with equatorial crossing times of 13:30 (ascending) and 1:30 (descending), uses a variety of spectra from infrared to microwave frequencies to retrieve H$_2$O in conditions with up to 70% cloud fraction. The Aura spacecraft lagged Aqua by about 14 minutes until the beginning of 2008, but due to the limb scanning geometry of MLS, its measurements only lagged those of AIRS by about 8 minutes. From May 15, 2008 the Aura spacecraft was maneuvered to co-locate MLS and CloudSat measurements, which reduced the lag between AIRS and MLS measurements to about 1 minute.

AIRS can retrieve temperature from the surface up to regions in the stratosphere ($\sim$10 hPa) and H$_2$O for $P \geq 250$ hPa with nominal vertical resolutions of $\sim$2-3 km for both parameters. MLS can retrieve H$_2$O from 0.1-316 hPa with a nominal vertical resolution of $\sim$2 km. The joint AIRS/MLS H$_2$O profiles developed in [LEI10] and AIRS temperature profiles, for the period of Aug, 2004 - Mar, 2010, are used for this analysis. Briefly, [LEI10] analyzed the AIRS and MLS averaging kernels (AK) (these are functions that describe the vertical sensitivity of passive remote sensing instruments to atmospheric trace gases) and confirmed that the AIRS and MLS have overlapping sensitivity to UT H$_2$O around $\sim$260 hPa; this was also shown through direct H$_2$O comparisons by [RLB07] and [FRW08]. Both instruments have vertical resolutions that range between $\sim$2-3 km. However, for pressures $P < 260$ hPa the AIRS sensitivity, diagnosed from the AK, drops dramatically while the MLS has strong sensitivity to H$_2$O throughout
the TTL region. Therefore, for $P < 260$ hPa a weighted mean was constructed between the AIRS and MLS profiles using the AK as weighting factors. Because MLS saturates often and AIRS has high sensitivity to H$_2$O for $P \geq 260$ hPa the AIRS data was used for these pressures. The data domain for this work will be from 1013 hPa - 10 hPa. We do not account for the differences in the lag periods before and after the May 15, 2008 spacecraft maneuver. We focus this analysis in the deep tropics between 8°S-8°N because it was found that the 100 hPa temperature signal associated with the QBO (for this data record) is strongest in this latitude band.

In order to compute $RH$, the AIRS temperature profiles are used throughout the entire data domain per the recommendation of [RLB07]. The saturation vapor pressure ($e_s$) over liquid and ice are computed using [MK05]. The liquid and ice $e_s$ are taken over temperature ranges of $T \geq 0$ °C and $T \leq -20$ °C respectively. $RH$ is just the ratio between the H$_2$O concentration and the saturation mixing ratio ($q_s$). For $-20 < T < 0$ °C $RH$ is a linear interpolation between the values for ice and liquid.

Quantifying the sampling of the data is necessary in order to identify any measurement biases. Figure 3.1 shows the joint sampling characteristic of AIRS and MLS. Figures 3.1a and 3.1b show the number of AIRS (H$_2$O and temperature) and MLS (H$_2$O) full soundings respectively for each 4° x 4° box over the latitude band of ~50S-50N. AIRS full soundings, in this study, are defined as profiles that have quality flags PGood = Psurf, i.e. profiles of PGood quality that reach the surface (see [te10a] for quality flag details). MLS full soundings are profiles (after a series of quality checks) with H$_2$O retrievals for pressures $P \leq 316$ hPa (see [te07a] for quality flag details). Figures 3.1c and 3.1d report the values in Figure 3.1a and 3.1b, respectively, as a percentage of the total number of soundings (of
any quality) in each grid box. One clear feature in Figure 3.1d is that the low sampling essentially follows the Inter-Tropical Convergence Zone (ITCZ), a region with frequent deep convection. Although MLS can sound through optically thick clouds, the large ice hydrometeors in deep convective cloud tops cause enough scattering to impede the soundings. In Figure 3.1c one observes that AIRS also has sounding limitations in the tropics, primarily in regions of deep convection and persistent stratiform clouds. This is simply a limitation of the IR instrument since these stratiform clouds are usually optically thick enough to prevent any possible sounding through them. Furthermore, the AIRS cloud clearing algorithm requires some clear-sky in its field-of-view in order to estimate the clear versus cloudy radiances [SBB06]. Stratiform clouds regularly have cloud fractions on the order of 80-100%, preventing AIRS from obtaining any useful information from the clear-sky radiances. Figures 3.1e and 3.1f show the annual mean cloud fractions reported by AIRS for its top and bottom cloud top pressures (CTP) (AIRS can report CTP for up to 2 cloud layers). One can see that regions of high (low) cloud fraction almost exactly overlay the regions of low (high) sampling in Figure 3.1c. Most of the clouds in Figure 3.1e are ice clouds with a peak in the CTP distribution around 200 hPa (not shown), while the clouds in Figure 3.1f primarily consist of water clouds with a peak in the CTP distribution around 650 hPa (not shown).

From Figure 3.1a and 3.1c one might gather that AIRS can provide a high yield of tropical soundings of H$_2$O. However, [LEI10] showed that AIRS loses sensitivity to H$_2$O at levels above the 250-260 hPa layer in the 40S-40N latitude band. Using just the MLS profiles for pressures as high as 316 hPa would lead to the sampling in Figure 3.1d. However, combining the AIRS and MLS profiles as described in [LEI10], which uses AIRS from 1013-261 hPa and merges the AIRS and MLS H$_2$O profiles from 215 hPa to the top of atmosphere, results in the
Figure 3.1: Statistics are for Aug, 2004-Mar, 2010; they are gridded in 4° x 4° boxes, from 50°S-50°N: (a) Number of full AIRS temperature and H2O soundings, (b) Number of full MLS H2O soundings down to 316 hPa, (c,d) same as (a,b) but reported as a percentage of the total # of soundings, (e,f) the mean AIRS cloud fractions for the reported top and bottom retrieved cloud layers, (g) percent of grid boxes with usable merged profiles, and (h) the percentage difference between the number of full soundings between AIRS and MLS and the number merged profiles (pixels for which AIRS has full soundings and MLS has profiles down to 215 hPa, rather than 316 hPa).
sampling shown in Figure 3.1g. Figure 3.1h shows the difference between using AIRS and MLS individually, i.e. the union of Figure 3.1c and 3.1d, versus the merged dataset. One finds that the sampling increases by as much as 30% in the tropics when merging the profiles.

It is noted that because both instruments have lower sampling in the tropics (due to high cloud cover), the subsequent H$_2$O climatologies are “dry biased” towards clear sky conditions. Nevertheless, as will be shown in Section 3.3.4.5, the temperature (AIRS) and H$_2$O (AIRS and MLS) profiles still capture the latent heating and free tropospheric moistening associated with deep convection. Also, although the sampling is still biased towards lower cloud fractions, there are still over $\sim 10^6$ equatorial soundings, in clear and cloudy conditions, that provide enough information to discern the interannual variability of temperature and H$_2$O due to the ENSO and QBO.

From the combined atmospheric profiles, an equatorial ($8^\circ$S-8°N and 180°W-180°E) record of H$_2$O, temperature, and $RH$, from 1013 hPa - 10 hPa, is derived. The record is averaged in $\sim 16$ day bins (15 days for January, 16 days for January-December, and 14 (15) days for December in a 365 (366) day year) following the A-Train repeat cycle, i.e. the number of days it takes the A-Train satellites to repeat a particular orbital track. This helps to enhance the sampling size and minimize bias caused by missing data. The spatial distribution of the interannual signals is computed to identify the regions where the ENSO and QBO impact the tropopause region.

### 3.2.2 Ocean Niño Indices

Although the data record is short there are six ENSO events between Aug, 2004-Mar, 2010. The ENSO events, identified by the Ocean Niño Indices (ONI)\cite{te10b},
are shown in Figure 3.2. The ONI quantify the interannual sea surface temperature anomalies (SSTA) in the TCP (Niño3.4 region), with positive and negative SSTA associated with El Niño and La Niña. Each ONI comes from a running mean of SSTA over three consecutive months (e.g. Dec, Jan, Feb (DJF); Jan, Feb, Mar (JFM)). Periods with three-month means exceeding $0.5^\circ K$ indicate significant ENSO events (green markers in Figure 3.2). Figure 3.2 also show the relevant ENSO composites for this study. Briefly, the 1:E+Q+ composite includes the two El Niño events surrounding DJF of 2005 and 2007, when the ENSO and QBO (westerly) are both in phase. The 2:E-Q- composite include the two La Niña events surrounding DJF of 2006 and 2008, when the ENSO and QBO (easterly) are in phase. 3:E-Q+ and 4:E+Q* represent single seasonal composites of the 2009 La Niña and 2010 El Niño events which serve to highlight periods when the ENSO and QBO are out of phase (technically 4:E+Q* corresponds to an easterly transition of the QBO, pulling the QBO out of phase of El Niño).

### 3.2.3 Quasi-Biennial Oscillation Indices

Figure 3.2 also shows the Quasi-Biennial Oscillation Indices (QBOI), which are derived from the 50 hPa zonal wind anomalies (zonally averaged over the equator; indices come from the National Centers for the Environmental Prediction (NCEP)/National Center for Atmospheric Research (NCAR) Climate Data Assimilation System (CDAS)). These are used to diagnose the amplitude and phase of the QBO because: 1) the onset of the QBO is near 50 hPa typically around boreal spring, with maximum amplitudes around boreal winter [BGD01] and 2) the 50 hPa wind anomalies are best correlated with tropopause temperatures since the zonal wind shear at this level is in thermal wind balance with temperature anomalies in the lowest parts of the stratosphere [RWG00]. Positive and negative
Figure 3.2: Time series of ONI (red) and QBOI (black). Green markers identify when SST anomalies exceed a magnitude of |0.5K|. Specific ENSO composites are identified by 1:E+Q+, 2:E-Q-, 3:E-Q+, and 4:E+Q* (see text for details). The ENSO and QBO are in phase from the beginning of the time record until approximately boreal spring of 2008. Beyond this time the ENSO and QBO fall out of phase until the end of the record.
indices correspond to westerly and easterly wind anomalies. A full QBO period consists of a stratospheric westerly and easterly anomaly pair corresponding to positive and negative temperature anomalies. There are approximately two full QBO periods within the A-Train record, with periods of $\sim 24$ and $\sim 30$ months from boreal spring of 2005 to boreal spring of 2007, and the boreal spring of 2007 to boreal winter of 2010 respectively. We briefly note that [HH01] use the 50-70 hPa wind shear as a QBO index rather than just the 50 hPa zonal winds, however, the choice of index does not impact our work as we produce similar results to theirs.

3.3 Results

3.3.1 Time Record Mean

In this section, the basic tropical structure of temperature and H$_2$O is quantified. The time record means, taken from 8$^\circ$S-8$^\circ$N in longitude bins of 4$^\circ$, are computed for temperature (Figure 3.3a), H$_2$O (Figure 3.3b), and relative humidity (Figure 3.3c) with the time record zonal mean removed at each pressure level and longitude bin. Temperature is reported as an absolute deviation from the time record zonal mean in Kelvin. Relative humidity is treated in the same manner but reported in percent (%). In order to show the H$_2$O structure through the entire vertical domain (1013 hPa - 10 hPa), the anomalies are computed as relative percent (%) deviations from the time record mean for each pressure level and longitude bin.

The quadrupole structure in temperature (Figure 3.3a), from $\sim 60$-400 hPa and 100$^\circ$E-60$^\circ$W, reveals the typical dipole behavior of convection in the TWP and Indian Ocean (labeled as IO in Figure 3.3) and subsidence in the Niño3.4
region (herein referred as the TCP). [FDD09] confirmed a similar temperature structure with the European Centre for Medium-Range Weather Forecasts (ECMWF) reanalyses (their Figure 6a). The region between 150-70 hPa (TTL as defined by [FDD09]), reveals strong cold (warm) anomalies above the free-tropospheric heating (cooling) in the TWP (TCP). As in previous studies, e.g. [HN07, SD01], this is the expected behavior that results from tropical convection, although these studies (and perhaps others) ascribe this temperature pattern to different aspects of convection. A peculiar warm anomaly feature resides between 500-700 hPa and ∼75°W-180°W, most likely due the latent heating of deep convective clouds, north of the equator, that persist throughout the year. Referring back to Figure 3.1e and 3.1f one finds, along the equator from the TCP to South America, a thin band of high cloud fractions. This corresponds to the AIRS CTP distribution peak between 600-700 hPa directly within the warm anomaly region over the TCP (not shown). The thin warm layer extends to the dateline then expands to the large warm anomalies seen throughout the free troposphere. The AIRS top layer cloud fractions in Figure 3.1e (top layer cloud distribution peaks around 200 hPa) show this transition between the TCP and TWP, revealing a zonal climatological signature of a transition from shallow convection to deep convection.

Over the land masses of central Africa (labeled as AFR in Figure 3.3, 0°-40°E) and northern South America (labeled as SA in Figure 3.3, centered around 60°W), there are particularly large warm anomalies due to land surface warming; the anomalies over central Africa are on the order of ∼10 K. The free troposphere above the land masses is cold even though deep convection frequents these areas [MW92]. Since the temperature anomalies are calculated as deviations from the zonal mean, the cold anomalies from ∼70°W-70°E are likely due to the TWP and Indian Ocean having a much warmer free troposphere than the rest of the
The H$_2$O time record mean (Figure 3.3b) shows a dipole structure between the TWP/Indian Ocean (moister, convective) and TCP (drier, subsidence) regions from the surface up to the lower half of the TTL. Local extrema of moist (dry) anomalies, centered around 450 hPa, exist over the TWP (TCP). The moist features over central Africa in Figure 3.3) and South America are probably due to the moistening of deep convection, but the near surface anomalies are drier since they are over land masses. Although cold anomalies exist over these land masses in the free troposphere, evaporative moistening seems to dominate these regions. One other feature to note is the slight positive H$_2$O anomalies of 2-9%, i.e. 0.1-1 ppmv, from 70-150 hPa over these land regions.

In Figure 3.3c, the TTL RH distribution is tightly coupled to TTL temperatures due to the exponential dependence of $q_s$ on temperature and the aridity of the stratosphere. RH, in the UT and down throughout the free-troposphere and boundary layer, however, correlates more strongly with H$_2$O. Also, in the TWP there is a thin layer near the surface of slightly negative RH anomalies due to the slightly stronger temperature anomalies at the surface. Slightly positive RH anomalies are observed in the TCP in the boundary layer, near 120$^\circ$W, due to colder than average boundary layer temperatures. The positive RH in the TWP TTL results from the TWP having the coldest temperatures in the deep tropics leading to persistent saturation/supersaturation most of the year (see Table 3.1).

### 3.3.2 Time Evolution of Temperature, H$_2$O, and RH

Figure 3.4 shows the zonal mean time evolution, between 8$^\circ$S-8$^\circ$N, of (a) temperature, (b) H$_2$O, and (c) RH. Each time series represents ~16 day means (see Section 3.2 for details) reported as anomalies from the time record zonal mean
Figure 3.3: Temporal (from Aug, 2004-Mar, 2010) and meridional (08°S-08°N) mean, pressure-longitude (4° longitude bins) cross sections of (a) temperature (K), (b) water vapor (reported as a % departure from the record mean H₂O concentration at each pressure level for each 4° longitude bin), and (c) relative humidity (%). Locations of the Indian Ocean (IO), tropical western Pacific (TWP), tropical central Pacific (TCP), Africa (AFR), and South America (SA) are identified.
at each pressure level. The temperature anomalies range between ±6 K, with the largest anomalies above the tropopause region. The lapse rate $-\delta T/\delta z$ (not shown), between the surface and ~300 hPa, varies between 4-8 K/km, where the temperature roughly follows a moist adiabat in the tropical mean; the lapse rate between 125-300 hPa approaches a dry adiabat with values around 8 K/km. The time record shows predominately warm free tropospheric temperatures, consistent with the results of Figure 3.3a, with interannual impacts evident in 2006 (cold temperatures the whole year). The tropical mean 100 hPa temperature is ~192 K, consistent with values derived from reanalyses in [RWG00] and rawinsondes from [ZGZ01a]. [TRK06] showed that the mean AIRS bias between the surface and 100 hPa only ranged between ±1 K with the largest magnitudes occurring at 100 hPa. However, comparisons of AIRS and Constellation Observing System for Meteorology Ionosphere and Climate (COSMIC) soundings (boreal summer and fall) show the AIRS systematic error for 100 hPa temperatures is closer to ±~0.5K. (Personal Communication with Dr. Sun Wong at JPL/Caltech).

This indicates that the observed temperature signal throughout the atmosphere is greater than the measurement accuracy which gives us confidence in the observed temperature features.

Transitioning to Figure 3.4b, the well-known ascending H$_2$O tape-recorder [MRH96] (and references therein) is observed with anomalies reaching ±~40% of 3.9±0.3 ppmv. The time-averaged H$_2$O mixing ratio (3.9 ppmv) is consistent with the values of the stratospheric entry level H$_2$O concentrations listed in Table 1 of [ZGZ01a]. The stated MLS accuracy at 100 hPa is ~8% [RLB07], so the observed H$_2$O anomalies are well outside the measurement uncertainty. The H$_2$O mixing ratios increase with altitude in the stratosphere due to methane oxidation (not shown). The 100 hPa H$_2$O data presented here are a weighted mean between AIRS and MLS H$_2$O. However, since the 100 hPa H$_2$O values are heavily weighted
Figure 3.4: Equatorial mean (08°S-08°N, 180°E-180°W) time-evolution of (a) temperature (K), (b) H₂O (%), and (c) RH (%) with the time record mean removed at each pressure level. Data spans between Aug, 2004-Mar, 2010. Abscissa is time and ordinate is pressure. The black dash-dotted horizontal line denotes the 100 hPa level. Red text at the right ordinate are the time record means for the specified pressure levels at the left ordinate. Herein all contour time record plots of pressure vs time will contain these mean values.
by MLS, due to AIRS virtually having no sensitivity at this level, the MLS accuracy estimate still holds (this holds for $P < 100$ hPa as well) [LEI10]. There are time periods when the tropospheric moist (dry) anomalies share the same sign as the moist (dry) anomalies associated with the H$_2$O tape-recorder (e.g. see moist anomalies directly after January, 2007 and dry anomalies in the latter half of 2007 in Figure 3.4b). Within the H$_2$O series, aspects of the tropospheric interannual signal of H$_2$O, e.g. large areas of negative anomalies (blue) starting in the second half of 2007 and ending around the first half of 2009, are also discernible. In the time mean, the free troposphere is predominately moist, most likely due to the strong convection in the TWP, South America, and Africa (see Figure 3.3b). Interannual variations are also observed in the stratospheric H$_2$O tape-recorder [RWR98, GZZ02].

The mean tropical surface layer $\text{RH}$ (Figure 3.4c) is $\sim 70\%$ (shown at the bottom right ordinate) and drops off rapidly throughout the free troposphere, though it is still primarily moist (See Figure 3.3c). $\text{RH}$ rises again in the TTL up to the tropopause due to the strong temperature effect on $q_s$ with very low H$_2$O concentrations. A reduction of $\text{RH}$ once again occurs in the stratosphere due to the aridity of the stratosphere and the increase of temperature due to ozone absorption (not shown).

3.3.3 Cold Point Seasonal Cycle

The scientific literature indicates that the tropical mean H$_2$O amount that enters the stratosphere is $\sim 3.8$ ppmv with a mean $q_s$ of $\sim 4.5$ ppmv (e.g. [ZGZ01a, FDD09]). Table 3.1 summarizes the seasonal and annual means of $q_v$, $q_s$, and $\text{RH}$ for the equatorial tropics (Trop), TCP, and TWP. Previous studies (e.g. [MRH96]) have indicated that to determine the correct $q_s$ and stratospheric entry
Table 3.1: Seasonal and annual mean values of H$_2$O ($\bar{q}_v$), temperature ($T$), saturation mixing ratio ($\bar{q}_s$), and relative humidity ($RH$) at the profile cold point. Note: cold point temperatures were at 96 hPa 99% of the time. These are computed for the equatorial tropics (08°S-08°N,180°E-180°W), TWP (08°S-08°N, 120°E-170°E), and TCP (08°S-08°N,120°W-170°W). $\bar{q}_v$, $T$, $\bar{q}_s$, and $RH$ are in units of ppmv, K, ppmv, and % respectively. Parenthetical values in the “Season” row are the number of profiles that went into the calculation for (Tropical,TCP,TWP) data for each season. $\bar{q}_v$ error estimates come from the MLS accuracy of 8% at 100 hPa. A temperature accuracy of ±1 K [TRK06] is assumed for all $\bar{q}_s$ and $RH$ values.
amount of H$_2$O, one needs to compute these quantities at the coldest temperatures in each profile. AIRS temperature profiles show that the cold point lies at 96 hPa (99% of the time), slightly higher in pressure than previous cold point pressure estimates of ~90 hPa (e.g. [FDD09]). Since AIRS has coarser resolution in the UTLS (~3 km) it is not surprising that the observed cold point occurs at a higher pressure. It is noted that [Des98] and [ZGZ01a] estimated the stratospheric entry H$_2$O concentration from the minimum of $q_s$ rather than the minimum temperature derivation of $q_s$. Both methods were exercised for this work and the results of Table 3.1 do not change within the AIRS temperature uncertainty estimates.

The seasonal means show very clear seasonal cycles over all regions with boreal winter and spring corresponding to the driest and coldest seasons. The tropical annual mean $q_s$ ($\bar{q}_s$) of 4.9±0.9 ppmv is well within the estimates of the values listed in Table 1 of [ZGZ01a]. This estimate for $\bar{q}_s$ is also consistent with the estimate in [Des98]. The relatively large uncertainty on $\bar{q}_s$ is due to using a more conservative systematic temperature error estimate of 1 K [TRK06]. However, assuming a systematic error of ~ |0.5| K (from the AIRS and COSMIC comparison) essentially reduces all error bars in $\bar{q}_s$ and $RH$ by about a factor of 2. The tropical annual mean H$_2$O concentration at the cold point ($\bar{q}_v$) of 3.8±0.3 ppmv is well within the estimates in Table 1 of [ZGZ01a]. Note that we do not account for mid-latitude sources of H$_2$O that can contribute to tropical stratospheric H$_2$O concentrations [Des98]. Therefore, our estimate of the entry level H$_2$O concentration will be high. The TWP is supersaturated (with respect to ice) almost year around due to this area having the coldest tropical temperatures. The results in Table 3.1 verify the seasonal and annual means are consistent with previous studies, giving us confidence in the AIRS and MLS data.
### 3.3.4 Interannual Variability

Sections 3.3.2 and 3.3.3 showed that AIRS and MLS do capture the expected behavior of temperature and humidity within the instrumental uncertainty. Interannual variations were also identified in the data record. The two primary interannual modes that modulate tropical temperatures and H$_2$O are the ENSO and QBO, thus the remainder of this work will focus on these two modes of tropical variability. Note: For the rest of this work the 100 hPa temperatures will be used as a cold point proxy to study the effects of the ENSO and QBO on the temperature and H$_2$O distribution around the tropopause.

#### 3.3.4.1 Zonal Equatorial Mean Interannual Time Series

To remove the record mean annual cycle, the mean is computed (for temperature, H$_2$O, and RH) over all years in bins of ~16 days (see Section 3.2 for details on bin construction). This is then subtracted from the same 16 day bins for each year; Equation (3.1) illustrates this more clearly. The interannual signal for parameter “X” is calculated in the following manner:

$$
\Delta X_i^j = X_i^j - \sum_j X_i^j
$$

for the $i$th time record bin (~16 days/bin) within the $j$th year. $\Delta T$ and $\Delta RH$, the interannual signals for temperature and $RH$, are computed using the above equation. $\Delta q$, the interannual signal for H$_2$O, is expressed as a percent deviation from the record mean annual cycle by dividing Equation (3.1) by $\sum_j X_i^j$; Figure 3.5 shows these quantities.

Figure 3.5a shows the tropopause region $\Delta T$ has peak-to-peak amplitudes (~1.5 K) that are about half the observed maximum tropopause region peak-to-peak temperature anomalies (~4 K) presented in Figure 3.4a. Boundary layer
Figure 3.5: Interannual Variability of the equatorial mean (180°E-180°W, 08°S-08°N) of (a) temperature ($\Delta T$), (b) H$_2$O ($\Delta q$), and (c) RH ($\Delta RH$), in units of T, %, and % respectively. Panel (d) is the Ocean Niño Index (ONI, red) and the quasi-biennial oscillation U50 index (QBOI, black). Green markers indicate ENSO events (see Sections 3.2.2 and 3.2.3 for explanation of indices).
and free tropospheric $\Delta T$ follow the ONI with warm (red colors in Figure 3.5a) and cold (blue colors in Figure 3.5a) periods corresponding to El Niño and La Niña events respectively, highlighting the dominance of ENSO on the zonal mean temperature structure. In some periods, e.g. around January 2007 during an El Niño, $\Delta T$ in the UTLS ($\sim$150-70hPa) has the same sign as the free tropospheric $\Delta T$, while other periods, e.g. boreal winter of 2009 during a La Niña, the free tropospheric and UTLS $\Delta T$ have opposite signs. This is because the zonal mean anomalies in the LS down to 100 hPa is mainly determined by the downward phase propagation of the QBO, while the anomalies in the free troposphere and the UT are primarily controlled by ENSO which does not have a regular cycle like the QBO.

The $\Delta q$ signal (Figure 3.5b) is about half ($\sim$0.5 ppmv) the $H_2O$ anomalies ($\sim$1.2 ppmv) shown in Figure 3.4b for the peak-to-peak amplitudes. The vertically propagating signal (in time), starting around 100 hPa (denoted by the black dashed line), represents the interannual variability of the $H_2O$ tape-recorder [RWR98, GZZ02]. The zonal mean $\Delta q$ is also dominated by the ENSO through the free troposphere, and up to the bottom of the TTL, with moist (red) and dry (blue) regions corresponding to El Niño and La Niña events. This indicates the influence of the lower tropospheric processes on the UTLS $H_2O$ distribution that is associated with ENSO events. Once again, there are periods (see DJF of 2005, 2006, 2007, and 2008) when the prevailing free tropospheric $\Delta q$ share the same sign as those in the UTLS. This is a result of the joint impact of the ENSO and QBO being in phase and simultaneously dehydrating TTL air more (less) effectively during La Niña (El Niño) years [ZGZ04]. This simultaneous ENSO and QBO impact will be discussed in more detail below. $\Delta RH$ (Figure 3.5c) in the ULTS almost exactly mimics $\Delta T$ with the largest anomalies in the TTL. However, moving further down into the free troposphere $\Delta RH$ signal is reduced
from UTLS values and does not correlate well with the ONI or QBOI. This is consistent with RH remaining constant with global changes to temperature and H$_2$O (e.g. [SJR05]).

Shown in Figure 3.6a are the 1013 hPa (green) and 100 hPa (blue) $\Delta T$ for the entire equatorial region. The 100 hPa time record shows the cross section of anomalies denoted by the black dash-dotted lines in Figure 3.5. ENSO events alone should produce an anti-correlation between the 1013 hPa and 100 hPa $\Delta T$ since warmer (colder) SST should lead to stronger (weaker) convection which would lead to colder (warmer) tropopause temperatures (e.g. [GRM01]). There seems to be some anti-correlation in certain periods of $\Delta T$ between 1013 hPa and 100 hPa, however, computing a lag correlation over the entire time record results in a low correlation coefficient (R) of -0.13. This value is not surprising as: 1) our data includes land (land surface temperatures do not follow the ONI), 2) oceanic boundary layer $\Delta T$ are largely governed by the ENSO, and 3) 100 hPa $\Delta T$ are primarily governed by the QBO.

Therefore, one might expect the 100 hPa $\Delta T$ to follow the QBOI rather than the ONI. A correlation coefficient of R=0.86 (for lag 0) results from correlating the QBOI with 100 hPa $\Delta T$ with a confidence interval beyond 95%, reaffirming the thermal wind balance between 50 hPa and the lower stratosphere. The descending QBO $\Delta T$ is a minimum around 100 hPa, consistent with previous observations of the QBO (e.g. [Wal73]). The 100 hPa anomalies are about 50% greater than those at 1013 hPa. This tropospheric amplification is consistent with [SWM05] and [GF08] where they find that UT temperature anomalies are amplified by a factor of $\sim 1.5$ from the surface.

Figure 3.6b shows the same 100 hPa $\Delta T$ along with 100 hPa $\Delta q$ (pink). $\Delta q$ follows $\Delta T$ closely though with a lag of about a half a month. The lag
Figure 3.6: Equatorial mean (180°E-180°W, 08°S-08°N) interannual variability of (a) temperature at 1013 hPa (green) and 100 hPa (blue), (b) temperature (blue) and H₂O (pink) at 100 hPa. ONI and QBOI are plotted for reference in (c).
correlation between $\Delta T$ and $\Delta q$ yields $R=0.74$ for lag $+1$ ($\sim 16$ days) (at lag 0 $R=0.71$). The lag correlation between $\Delta q$ and the QBOI gives $R = 0.74$ (at lag $+4$) for lag of about 2.0 months ($R=0.60$ for lag 0). The cause of the difference in lag correlation between $\Delta T$ and $\Delta q$ and the QBOI and $\Delta q$ is unknown. The low correlation between $\Delta q$ and the ONI of $R=0.50$ (lag 0) affirms that the 100 hPa $\Delta T$ is predominately determined by the QBO. At 121 hPa, the correlation between $\Delta T$ and $\Delta q$ yields $R=0.78$ for lag 0, suggesting the $\sim 0.5$ month lag at 100 hPa is likely a manifestation of MLS averaging ("smearing") features of the H$_2$O tape-recorder from higher altitudes; MLS has a nominal vertical resolution of $\sim 2$ km which may not be able to resolve the details of the cold point tropopause (although other transport processes may play a role in the lag as well). The 100 hPa correlations between $\Delta T$ and $\Delta q$, as well as $\Delta q$ and QBOI, increases to $R=0.81$ and $R=0.80$ respectively when excluding data beyond Jan. 2008, which corresponds to when the ENSO and QBO begin to fall out of phase. This indicates possible different zonal impacts on H$_2$O depending on the relative phase of the ENSO and QBO. We note that doing correlations with time bins of $\sim 16$ days, on the one hand, yields more statistically significant time means, however, the coarseness of the time resolution prevents any probing of processes on scales finer than about a month. Thus, these lag correlations should only be seen as gross indications of relationships between remotely sensed parameters. Nevertheless, these correlations are statistically significant and confirm the strong influence of the QBO on the zonal temperature and H$_2$O structure, especially when the ENSO and QBO are in phase.

Therefore, the zonal mean structure of tropopause temperature and H$_2$O primarily follow the QBO. However, this mean structure masks out the zonally varying ENSO signal seen, for example, by [RWG00] and [GRM01]. The ENSO and QBO are approximately in phase from Aug, 2004 through boreal summer of
2008 (see Figure 3.3). After boreal summer of 2008 the ENSO and QBO fall out of phase. Some interesting questions to ask are: 1) how does 100 hPa $\Delta T$ depend on the phase of both interannual modes and 2) how is the $H_2O$ distribution in the UTLS related to the relative phase of the ENSO and QBO. These questions will be addressed in more detail in the sections below.

### 3.3.4.2 Zonal Structure of the ENSO and QBO at 100 hPa

In Section 3.3.4.1 the equatorial mean time record was computed ($8^\circ S$-$8^\circ N$, $180^\circ W$-$180^\circ E$). It was found that the QBO dominates the equatorial zonal mean $\Delta T$ and $\Delta q$ structure at 100 hPa. However, since the time record in Figure 3.5 and 3.6 are tropical means, it is difficult to identify any ENSO signature at 100 hPa. Thus, the longitudinal variations are investigated to explore the regional impacts of the ENSO and QBO in the tropopause region.

Figures 3.7a and 3.7b show longitude Hovmöller diagrams ($8^\circ S$-$8^\circ N$) of $\Delta T$ and $\Delta q$ respectively at 100 hPa. The 100 hPa $\Delta T$ (Figure 3.7a) varies coherently with the phase of the QBOI indicating the dominating impact of the QBO on 100 hPa temperatures. However, upon closer inspection of 1:E+Q+ and 2:E-Q- (periods when the QBO and ENSO are approximately in phase) some interesting features are observed. The TWP anomalies have the same sign as the coherent QBO $\Delta T$ and exhibits warmer (colder) anomalies than other longitudes during El Niño (La Niña) events. This is consistent with larger (smaller) dehydration volumes occurring during La Niña (El Niño) seasons [ZGZ04]. However, in the TCP, strands of opposite signed anomalies exist. ENSO events that produce anomalies opposite to the prevailing QBO signal are circled with a solid black oval.

From the boreal summer of 2008 to the first quarter of 2010 the ONI and
Figure 3.7: Longitude Hovmöller plots of (a) $\Delta T$ and (b) $\Delta q$, meridionally averaged in the $8^\circ$S-$8^\circ$N band, at 100 hPa (longitude bins are $4^\circ$). ONI and QBO are plotted for reference in (c). The TWP and TCP are also marked out by arrows. Ovals mark breaks in the zonal symmetry of the QBO (dashed oval for the TWP). Figure 3.2 is inserted in panel (c) for reference.
QBOI are out of phase. The oscillating QBO signal produces a prevailing warm $\Delta T$, especially between the boreal summers of 2008 and 2009 (composite 3:E-Q+). The ENSO and QBO temperature signatures are in phase in the TCP and the TWP experiences 100 hPa anomalies that are counter to the prevailing QBO signal. This is seen in the small blue bands circled with a dashed black oval. Figure 3.7b shows the strong impact of the QBO on the zonal 100 hPa $\Delta q$ distribution. Co-located with the regional temperatures circled in Figure 3.7a are areas of positive (negative) $\Delta q$ that correspond to positive (negative) $\Delta T$. However, the spatial coherence is weaker than with $\Delta T$.

To statistically quantify the zonal breaks in $\Delta T$ and $\Delta q$ at 100 hPa, the time record for the TCP (Figure 3.8a) and TWP (Figure 3.8b) are computed. Blue and pink patches indicate when $\Delta T$ and $\Delta q$, respectively, are statistically significant from 0 at the 95% confidence level; purple patches indicate when both anomalies are statistically significant.

The 100 hPa $\Delta T$ in the TWP (blue curve in Figure 3.8b) show relatively large amplitudes during the years when the ONI and QBOI (shown in Figure 3.8c) are in phase, with $\Delta T$ that range between -2.6 to +2.0 K (between the end of 2004 to the boreal summer of 2008), with statistical significant at the 95% level much of this period. $\Delta q$ is also statistically significant from 0 with anomalies that range between $\pm 0.8$ ppmv. During this period, the TCP experiences weaker anomalies with $\Delta T$ and $\Delta q$ that range between -1.8 to +1.3 K and $\pm 0.5$ ppmv respectively, with many oscillations around 0. When the QBOI and ONI fall out of phase (after the boreal summer of 2008), the TWP $\Delta T$ and $\Delta q$ falls off to smaller magnitudes between -0.8 to +1.8 K and $\pm 0.4$ ppmv with frequent oscillations around 0. The TCP now experiences statically significant anomalies (at the 95% level), for longer periods of time, with $\Delta T$ and $\Delta q$ ranging between -2.3 to +2.5
K and -0.4 to 0.5 ppmv respectively. The TWP $\Delta q$, as with the equatorial mean case, correlates well with $\Delta T$ giving R=0.77 (for lag +1), while the TCP shows a lower correlation of R=0.62 (0 lag). This is consistent with previous work identifying that $\Delta q$ is a function of the TWP CPT (e.g. [HG01]).

### 3.3.4.3 100 hPa Composite Maps of $\Delta T$ and $\Delta q$

To investigate further this apparent zonal break in the QBO signal, the structures of the 100 hPa $\Delta T$ (Figure 3.9a) and $\Delta q$ (Figure 3.9b) are investigated for boreal winter (DJF) in the context of the composites presented in Figure 3.2: 1:E+Q+, 2:E-Q-, 3:E-Q+, and 4:E+Q* (see Section 3.2.2 for composite details). For 1:E+Q+, the QBO induced, zonally symmetric, warm $\Delta T$ shows a break of reduced $\Delta T$ centered around 120°W-170°W (dashed circled region). This also occurs in 2:E-Q- with the positive $\Delta T$ induced by ENSO breaking the zonal symmetry of the negative QBO $\Delta T$. In both cases the ENSO and QBO are in phase and the observed $\Delta T$ around the zonal break forms a symmetrical dumbbell shaped pattern in the TCP since the zonal break has a maximum centered around 140W. These zonal asymmetries can also be interpreted as the QBO restricting (during the westerly regime, 1:E+Q+) or enhancing (during the easterly regime, 2:E-Q-) convection in the TWP [CHM98, CMH02].

Transitioning to periods when the ENSO and QBO are out of phase, 3:E-Q+ shows a break in zonal $\Delta T$ that now occurs over the TWP (dashed circled region) with the TCP taking on zonal anomalies of the same sign as the prevailing westerly phase of the QBO. During this period the La Niña in itself should increase the dehydration potential of the TWP (e.g. [ZGZ04]) via enhanced convection, however, the QBO westerly phase induces subsidence in the tropopause region, increasing tropopause temperatures, thus inhibiting convection from penetrating.
Figure 3.8: 100 hPa $\Delta T$ (blue curve) and $\Delta q$ (pink curve) for the TCP (panel (a)) and the TWP (panel (b)). Blue patches correspond to time periods when $\Delta T$ signals are statistically significant at the 95% level or better. Pink patches quantify the same statistics for $\Delta q$. Periods with purple patches correspond to instances when $\Delta T$ and $\Delta q$ are simultaneously statistically significant at the 95% level or better. Panel (c) shows the ONI and QBOI for reference.
Figure 3.9: 4° x 4° maps of 100 hPa (a) $\Delta T$ and (b) $\Delta q$ for composite events 1:E+Q+ (row 1), 2:E-Q- (row 2), 3:E-Q+ (row 3), and 4:E+Q* (row 4) discussed in Section 3.2.2. The "*" symbol indicates a nearly 0 QBOI. Dashed lines mark the 8°S-8°N band and the approximate meridional extent of the QBO signal in temperature. Dashed ovals mark locations of anomaly cancellations.

Figure 3.10: Same as Figure 3.9 but for a composite from the months of Aug-Oct (2008) when the ONI is nearly zero (see Figure 3.2)
deeper into the tropopause region [CHM98, CMH02].

The 4:E+Q* ∆T pattern appears, in contrast to the other ENSO events previously discussed, because the QBO is transitioning from the westerly to easterly phase, which takes place around boreal spring (see QBOI in Figure 3.2). Thus, the 4:E+Q* ∆T map shows the approximate temperature patterns one would expect with little influence from the QBO. The 4:E+Q* ∆T exhibits temperature patterns qualitatively consistent with the ENSO temperature patterns shown in Figure 6b of [KSR01]. The narrow band of warm anomalies in the TWP are enveloped by symmetrical cold anomalies around the equator in the TCP. These tropopause patterns are consistent with Rossby and Kelvin circulations induced by tropical equatorial heating (e.g. [Gil80, HH98]). Figure 3.10 shows, in contrast to 4:E+Q*, a three month mean (Aug, Sep, Oct) in 2008 when the ONI is near zero and the QBO reaches a local maximum. The zonal symmetry in ∆T is robust in the tropics with very little evidence of an ENSO signature on tropopause temperatures. The weaker convection during the fall season and weak ENSO allow the QBO signal to manifest as a zonally symmetric feature in temperature and H\textsubscript{2}O. Conceptually, the final signal in ∆T resembles a superposition of the QBO signal shown in Figure 3.10 and the ENSO signal in 4:E+Q* (Figure 3.9, 4th panel down).

The ∆q signals of 1:E+Q+ and 2:E-Q- show dominant prevailing signals that are in phase with the ONI and QBOI with 1:E+Q+ and 2:E-Q- showing positive and negative H\textsubscript{2}O anomalies respectively. This is a consequence of the ENSO and QBO being in phase such that the zonally asymmetric ENSO feature does not show up strongly. Although the ENSO is responsible for the anomaly reduction in the TCP, the anomaly enhancement in the TWP dominates the zonal ∆q distribution, consistent with Figures 3.6 and 3.8. However, 3:E-Q+ and 4:E+Q*
both show the TCP with anomalies of the opposite sign from the rest of the tropics. In these cases, $\Delta q$ is similar to $\Delta T$ in that a colder (warmer) tropopause corresponds to a drier (wetter) tropopause. This contrast in zonal $\Delta q$ for 3:E-Q+ and 4:E+Q* suggests that when the ENSO and QBO are out of phase, the TCP may play a more prominent role in regulating the tropopause region $H_2O$ distribution, though the source/sink of this $H_2O$ cannot be determined from the composites.

The resulting patterns of 100 hPa $\Delta T$ are interesting and have a simple explanation. During La Niña years (composite 2:E-Q- in Figure 3.9), deep convection is particularly strong in the TWP. This is associated with large positive temperature anomalies in the free troposphere due to convective heating (see Figure 3.3a). Associated with this strong heating is a thinner layer of anomalously cold temperatures around the tropopause region. In the TCP a warmer tropopause region is associated with colder free tropospheric temperatures. If the ENSO and QBO are in phase, then the cold ENSO anomalies will be in phase with the zonally symmetric cold QBO anomalies in the TWP, leading to an enhancement of cold anomalies, thus leading to greater dehydration in $H_2O$ [ZGZ04]. These same cold anomalies will work against the warm anomalies that are seen in the TCP tropopause temperatures, resulting in reduced anomalies. During the El Niño years that are in phase with the QBO (composite 1:E+Q+ in Figure 3.9), one expects the TCP to have colder tropopause temperatures. If the ENSO and QBO are in phase, once again, the zonally symmetric QBO signal, now positive (warm), will counteract the cold TCP tropopause temperatures while enhancing the warm tropopause temperatures in the TWP, leading to moistening/reduced dehydration in the TWP [ZGZ04]. In these cases, the zonal structure of $H_2O$ is primarily driven by the TWP CPT [HG01].
During the periods when the ENSO and QBO are out of phase (3:E-Q+ and 4:E+Q* in Figure 3.9), the westerly (easterly) anomalies will enhance the TCP warm (cold) tropopause anomalies associated with La Niña (El Niño) events, while reducing the cold (warm) anomalies in the TWP. Revisiting 3:E-Q+, we find that a prominent H₂O maximum in TCP that is not present when the ENSO and QBO are in phase. This maximum is consistent with the La Niña warm top over the TCP being in phase with the warm QBO westerly, although the source of the H₂O maximum is not known.

### 3.3.4.4 Isolating the ENSO Signature

In Section 3.3.4.2, it was found that the ENSO and QBO either enhance or reduce temperature and H₂O anomalies at 100 hPa depending on the phase and location of observation. The anomalies are magnified in the TWP when the ENSO and QBO are in phase, leading to reduced anomalies in the TCP. When the ENSO and QBO are out of phase, the TCP becomes the region of enhanced anomalies while the TWP anomalies counter the prevailing QBO signal. Although the statistics seem to support this hypothesis, the ENSO effect still has not been quantified.

In order to isolate the ENSO signal, the approximate zonal symmetry nature of the QBO is exploited. This is done by subtracting the tropical zonal mean time record (Figure 3.5) from the corresponding TCP and TWP time record.

In the TWP, the correlation between \( \Delta T_E \) and \( \Delta q_E \) (\( \Delta T \) and \( \Delta q \) with their zonal mean subtracted out) at 100 hPa only yields \( R=0.48 \). This is in contrast to the high correlation \( (R=0.77, \text{lag } +1) \) for the case when the zonal mean is included. Within the TCP, the 100 hPa correlation between \( \Delta T_E \) and \( \Delta q_E \) yields \( R=0.82 \) (lag 0). This high value suggests that the local temperatures in the TCP do have a strong influence on the local H₂O distribution. During El
Niño (La Niña) years, the TCP tropopause region cools (warms) leading to a dryer (moister) tropopause region. Recall, a weaker correlation was computed between $\Delta T$ and $\Delta q$ with the zonal mean included ($R=0.62$, lag 0). This, in part, may explain why when the ENSO and QBO are out of phase the TCP $\text{H}_2\text{O}$ maxima shows up (e.g. 3:E-Q+). When the QBO and ENSO are out of phase, the ENSO impact on tropopause $\Delta T$ and $\Delta q$ is now supported by the QBO.

### 3.3.4.5 Horizontal and Vertical structure of the ENSO and QBO

We now investigate the vertical and horizontal spatial distribution (Figure 3.11) of the interannual variability of temperature and $\text{H}_2\text{O}$ to corroborate the results shown in Sections 3.3.4.2-3.3.4.4. The same time period composites as shown in Figure 3.9 are selected for analyses, however, in order to highlight the impacts of ENSO, specifically El Niño, we compute the difference $[1:E+Q+] - [2:E-Q-]$ (herein EL01, brackets inserted for readability) and $[4:E+Q^*] - [3:E-Q+]$ (herein EL02).

The EL01 $\Delta T$ composite (Figure 3.11a) shows a strong thick band of warm QBO anomalies from 50 hPa down into the TTL. The TWP TTL shows especially warm anomalies as compared to the rest of the tropics contrasting the thin vertical band of cold anomalies over the TCP TTL. The tropical troposphere is also warm. As previously discussed (section 3.3.4.2), the ENSO dominates the free tropospheric and boundary layer zonal mean $\Delta T$. Subtracting the zonal mean $\Delta T$ produces $\Delta T_E$ (Figure 3.11c), revealing the asymmetrical ENSO anomalies (with possible other effects as well; e.g. tropospheric biennial oscillations which impacts aspects of the rainfall associated with the Indian Monsoon [CL00]). A quadrupole temperature anomaly structure is revealed in $\Delta T_E$, with a cold top over a warm free troposphere in the TCP and colder free troposphere in the TWP with a warm
Figure 3.11: Vertical cross sections of: (a,b) $\Delta T$, (c,d) $\Delta T$ with the zonal mean removed ($\Delta T_E$), (c,f) $\Delta q$, and (g,h) $\Delta q$ with the zonal mean removed ($\Delta q_E$) for composite differences between $[1:E+Q+] - [2:E-Q-]$ (EL01, row 1) and $[4:E+Q^*] - [3:E-Q+]$ (EL02, row 2) (composite events shown in Figure 3.9). Ordinate is pressure in hPa and abscissa longitude (4° bins); meridional mean is taken from 8°S-8°N. Overlaid solid black horizontal line marks 100 hPa. Overlaid blue curve is the interannual mean rain rate anomalies ($\Delta rr$), in mm/hr, derived from TRMM with ordinate axes on the right; dashed-dotted line marks zero $\Delta rr$. Brackets for composite labels serve for readability purposes only.
A quadrupole pattern in $\Delta q_E$ is roughly co-located with the $\Delta T_E$ structure. However, there are a few differences in the $\text{H}_2\text{O}$ structure. First, the tropospheric anomalies in $\Delta q_E$ extend into the bottom of the TTL where one finds $\Delta T_E$ with opposite sign to $\Delta q_E$. Secondly, the layer of TTL $\Delta q_E$ anomalies primarily lie above 100 hPa in the TCP, whereas $\Delta T_E$ is approximately symmetric around 100 hPa. Thus, although the ENSO impacts on temperature are symmetric around the tropopause, its impact on $\text{H}_2\text{O}$ is skewed into the stratospheric region.

The $\Delta T_E$ signal for EL01 clearly shows the cold TCP TTL is playing a role in reducing the warm QBO anomalies while the warm TWP TTL (Figure 3.11c) enhances the QBO signal. Associated with these warm anomalies is hydration of the TTL region over the TWP (in a La Niña case, i.e. $[2:E-Q-] - [1:E+Q+]$, there would be enhanced dehydration, highlighting composite $2:E-Q-$). Upon further inspection, one finds that the mean $\Delta q$, i.e. the mean $\Delta q$ difference $[1:E+Q+] - [2:E-Q-]$, in the TCP is $\sim 2\%$ ($\sim 0.1$ ppmv), while while the TWP mean is $\sim 12\%$ ($\sim 0.4$ ppmv). EL02 shows that deep convection in the TCP leads to cold TTL $\Delta T_E$ that enhances the QBO cold anomalies. The overlaid (light blue curves in Figure all panels in Figure 3.11) interannual rain rate anomalies ($\Delta rr$, in mm/hr), computed from Tropical Rainfall Measuring Mission [IKM00] data corroborates that convection follows the ENSO cycle; high $\Delta rr$ follows the tropospheric moistening. For EL02, the TCP cold anomalies associated with deep convection are now allowed to constructively work with the QBO to dehydrate TTL air over the TCP, or in the case of $3:E-Q+$ (a La Niña), moisten the TTL TCP. In the case of $[3:E-Q+] - [4:E+Q^*]$ (highlighting La Niña), the mean TCP and TWP differences in $\Delta q$ are $\sim +8\%$ ($\sim +0.3$ ppmv) and $\sim -13\%$ ($\sim -0.4$ ppmv) respectively, with the TCP TTL anomalies now switching signs. Although $3:E-Q+$ shows a moist anomaly over the TCP, presumably from the QBO westerly anomaly and La Niña warm top having the same sign, the source of this $\text{H}_2\text{O}$
cannot be determined from this analysis.

3.4 Discussion

We have utilized a merged temperature (AIRS) and H$_2$O (AIRS and MLS) dataset [LEI10] to create a full-atmospheric time record of temperature and H$_2$O quantifying both the annual cycle and interannual variability of these quantities. This dataset captures, in a self-consistent way, the connection between the troposphere and stratosphere for temperature and H$_2$O, something that previous observation records could not capture, especially for H$_2$O. From the seasonal and annual cycle, we estimated the mean H$_2$O concentrations, $q_s$, and temperature as a function of various locations. Our estimates, within instrument uncertainty, are consistent with previous estimates of these values and indicate that the AIRS and MLS combined dataset does capture the zonal structure of the tropics well.

We were also able to identify the interannual signals of the ENSO and QBO, showing that the boundary layer and free tropospheric interannual variations of convection can cancel or enhance tropopause region temperature anomalies depending on the location and phase of the ENSO and QBO. When the ENSO and QBO are in phase the TWP temperature anomalies are enhanced while the TCP experiences anomalies counter the prevailing QBO anomalies. However, when these interannual modes are out of phase, the TCP experiences the anomaly enhancements with the TWP showing weaker anomalies. The hypothesis that these anomaly enhancements and cancellations are a result of changes to the Walker Circulation, which lead to a migration of convection between the TCP and TWP, is further supported by the migration of free tropospheric H$_2$O anomalies and the local strengthening or weakening of regional rain rates. However, we found that when the ENSO and QBO are in phase the ENSO impact on the zonal H$_2$O
distribution is masked by the QBO, though we do identify enhanced (weakened) dehydration in the TWP and zonal mean H$_2$O, consistent with previous work [ZGZ04]. On the other hand, when the ENSO and QBO are not in phase, the ENSO impact on H$_2$O is not only highlighted in the TWP but also the TCP. This is consistent with the enhanced warming observed in 3:E-Q+ over the TCP. In addition, the high correlation between $\Delta T_E$ and $\Delta q_E$ suggests that H$_2$O in the TCP is locally regulated. However, more work needs to be done to investigate H$_2$O transport processes in the TCP, particularly when the ENSO and QBO are out of phase. We note that our results particularly apply to boreal winter interactions of the ENSO and QBO. [ZGZ04] indicated that not only does the relative phase of the ENSO and QBO matter but also the time of year they interact. However, they do indicate that the interaction between the ENSO and QBO are at a maximum during boreal winter since tropical convection and tropopause height are also at a maximum during this time of year. Our findings are consistent with their results.

Figure 3.11 also highlights that while there are quadrupole patterns of anomalies in temperature ($\Delta T_E$) and H$_2$O ($\Delta q_E$), the spatial extent of the anomalies are different. In temperature, the anomalies due to convection in the troposphere only reach $\sim$ 200 hPa, roughly the location of neutral buoyancy in the tropics, however, the H$_2$O anomalies reach $\sim$120 hPa. In the TTL, the temperature anomalies are roughly symmetric about the tropopause but H$_2$O shows particularly strong anomalies in the LS especially over the TCP. Because our calculations only capture an Eulerian framework of the temperature and H$_2$O distribution, it is difficult to determine the possible transport pathways that determine the regional H$_2$O structure. For example, although 3:E-Q+ shows enhanced warming and moistening over the TCP, the source of this moistening cannot be determined from our analysis. Although, since 3:E-Q+ was during a La Niña, convective in-
jection from the TCP is not likely the source of the observed stratospheric moist anomaly. A longer time series would aid in corroborating our findings. It is also important to note that we cannot, from the observations, determine whether the zonal asymmetries are a linear superposition of the ENSO and QBO impact on the tropopause or a coupling between the two interannual modes. There is evidence, however, that the ENSO can alter planetary waves and their propagation that may modulate the magnitude and phase of the QBO [Dun97].

What has not been addressed in this work is the role of clouds in the TTL. It is well known that the TWP has a high frequency of occurrence of thin cirrus (e.g. [CAT04]) and is frequently supersaturated (see Table 3.1). [HG01] showed in their simulations that the TWP could in principle dehydrate air to the H$_2$O concentrations consistent with lower stratospheric values from frequent horizontal passes of air parcels through this cold region heavily populated by TTL cirrus. Furthermore, the presence of TTL cirrus modulates the level of zero radiative heating (LZRH), a metric of the radiative balance in the tropical atmosphere [GFM04, CLP05]. The radiative balance determines the location where tropical upwelling departs from being primarily convective to becoming radiatively driven. Thus, the tropical distribution of TTL cirrus is important for not only quantifying the H$_2$O distribution but also the time scale of its transport.

Our findings do have some implications. From a regional atmospheric process perspective, we found that the TCP can play a significant role in impacting the zonal H$_2$O distribution, particularly when the ENSO and QBO are out of phase. Since it is still not known if there is a preferred tropical longitude for where H$_2$O enters the stratosphere (e.g. [Des98]), the TCP, in principle, can impact the stratospheric H$_2$O tape-recorder. The presence of the 3:E-Q+ moist anomaly over the TCP stratosphere is evidence of this impact. From a modeling
standpoint, any model would need to capture the relative phase of the ENSO and QBO correctly, to simulate the interannual impacts on the zonal $\text{H}_2\text{O}$ distribution correctly. Furthermore, if high cloud amount is found to be crucial in quantifying UTLS $\text{H}_2\text{O}$, models would need to capture the correct horizontal and vertical distribution of clouds.
CHAPTER 4

Interannual Variability of Tropical Clouds

“Do you wish to rise? Begin by descending. You plan a tower that will pierce the clouds? Lay first the foundation of humility.” – Saint Augustine

4.1 Preamble

In addition to the tropical tropopause layer (TTL) temperature and H$_2$O structure, the tropical high cloud distribution is also impacted by interannual modes of variability. Symmetrical equatorial heating in the equatorial belt leads to excitation of Kelvin and Rossby waves which impact not only the boundary layer/free tropospheric temperature structure but also the TTL temperature structure via deep convection [Gil80]. The sign of the dipole TTL temperature anomalies between the TWP and TCP are modulated by the ENSO cycle. Recent work has provided evidence that these wave induced temperature anomalies also impact the tropical thin cirrus distribution [VWF10]. The ENSO cycles also cause shifts in the Walker Circulation which manifests itself as a migration of deep convection and, therefore, precipitation between the TCP and TWP [Phi90, LEG11]. Furthermore, previous work found evidence that the quasi-biennial oscillation (QBO) [Ree65b, Ree65a, Ham84, Dun85, BGD01], the other significant tropical interannual mode, may impact the distribution of deep convective clouds.
In Chapter 3 we observationally quantified the ENSO and QBO impact on the TTL temperature and H$_2$O distribution. Because both these modes have such distinct and significant impacts on tropical thermodynamical structure, the distribution of cloud amount, in principle, should also be impacted. The changes in the cloud distribution are important to quantify because they significantly affect the tropical radiative balance which impacts the transport pathways of trace gases into the stratosphere [CLP05]. Thus, this work will serve to answer the following questions: how do sea surface temperature changes associated with ENSO change the total cloud frequency of occurrence in the tropics, and does the QBO have any impacts on the tropical cloud structure?

4.2 Data and Methodology

4.2.1 Cloud Profiles

The cloud profiles come from two active instruments: 1) the Cloud Profile Radar (CPR) on board CloudSat [SVB02] and 2) the Cloud-Aerosol Lidar with Orthogonal Polarization (CALIOP) on board the Cloud-Aerosol Lidar and Infrared Pathfinder Satellite Observation (CALIPSO) [WPM03] spacecraft. The CPR operates in the microwave at 94 GHz and CALIOP has two active channels of 532 nm and 1064 nm. The CPR actively profiles clouds at horizontal and vertical resolutions of $\sim$1 km and $\sim$500 meters respectively, while the CALIOP can resolve down to 333 meters in the horizontal and 30-60 meters in the vertical. Both these instruments operate at disparate parts of the electromagnetic spectrum to provide the capability to profile clouds of varying optical thickness. Because the CPR wavelength regime is in the microwave, it is able to actively sense through...
entire cloud columns with large hydrometeors (strictly, this isn’t true for heavily precipitating parts of a cloud). However, because the CPR wavelength is so large it is unable to detect clouds with very small ice-crystals such as tropical tropopause layer (TTL) cirrus, sometimes termed “sub-visual” cirrus. This is where the CALIOP becomes useful. Combining both of these instruments provides an unprecedented capability to observe the global distribution of clouds while preserving their detailed vertical distribution.

From these two instruments the following datasets have been derived to easily interpret the observations. This analysis uses the CloudSat: 1) 2B-GEOPROF [MZV09], 2) 2B-GEOPROF-LIDAR (a combined Cloudsat and CALIPSO product) [MZV09], and 3) 2B-CLDCLASS [SW08]. Cloud data in the latitude band 10°S-10°N are used to restrict the data to the deep tropics. The 2B-GEOPROF-LIDAR product reports cloud layers determined by both CloudSat and CALIPSO. The 2B-GEOPROF-LIDAR puts both the CloudSat and CALIPSO data on a height grid with a vertical gradation of ~240 meters. The horizontal spacing is ~1 km. The data spans from Aug, 2006 - Feb, 2011.

4.2.2 Cloud Classification

The CloudSat clouds are classified according to the CloudSat 2B-CLDCLASS product. Since there is no equivalent for CALIPSO, to classify the clouds observed by the CALIOP, the vertical cloud structure is partitioned into two regimes: 1) clouds with base $Z_B \geq 7$km and 2) $Z_B < 7$km. We classify clouds above 7 km as thin cirrus, while clouds below 7 km are classified as “other.” The “other” class clouds may contain some cirrus but primarily contains optically thin boundary layer (liquid) and mid-level (mix-phased) clouds. The 7 km threshold is based on the Cloudsat 2B-CLDCLASS criteria for “High Clouds” (see [te07b] Table 1).
The are times when the 2B-GEOPROF identifies multiple cloud types within a single 2B-GEOPROF-LIDAR layer. To avoid the ambiguity of assigning multiple cloud types to a single 2B-GEOPROF-LIDAR layer, we default to the 2B-GEOPROF layer definitions which preserves the CloudSat classification scheme; this is done if the cloud detection confidence is $\geq 20$ [te07b]. Finally, for clouds that are identified by both the CPR and CALIOP, we choose the CALIOP assignment for cloud tops and the CPR for cloud bases. This is done because the visible pulses from the CALIOP will more likely be attenuated at the cloud top, whereas the microwave is more able to penetrate to the base of the cloud.

### 4.2.3 Ancillary Datasets

In addition to the Cloudsat and CALIPSO combined products, we also analyze rain rate data from the Tropical Rainfall Measuring Mission (TRMM) [IKM00] and cold-point tropopause data (Data provided by Dr. Sean Davis of NOAA ESRL Chemical Science Division) derived from the Modern Era Retrospective-analysis for Research and Applications (MERRA) [MER08, RSG10] reanalyses. In order to quantify the phase of the ENSO and QBO, we utilize the same Ocean Niño indices [te10b] and the QBO indices (QBOI) described in Sections 3.2.2 and 3.2.3 (record now spanning Aug, 2006 - Feb, 2011). The five ENSO events that will be analyzed are the El Niño seasons of boreal winter (DJF) 2006-2007 (1:E+Q+) and 2009-2010 (4:E+Q*), and the La Niña seasons of boreal winter 2007-2008 (2:E-Q-), 2008-2009 (3:E-Q+), and 2010-2011 (5:E-Q+). Events 1:E+Q+ and 2:E-Q- correspond to when the ENSO and QBO are in phase while 3:E-Q+, 4:E+Q*, and 5:E-Q+ mark when they are out of phase.
4.3 Results

4.3.1 Tropical Cloud Distribution

4.3.1.1 Time Record Mean

In this section we compute the cloud frequency of occurrence (CFO) over the entire time period of the data as a function of height and longitude within the 10°S-10°N latitude band. In altitude, the data are spaced in 240 meter bins to mimic the approximate vertical spacing of the CloudSat 2B-CLDCLASS and 2B-GEOPROF products. The data is also binned in 4° x 4° grid boxes to better capture the climatology of CFO.

Figure 4.1a shows the height-longitude time-record mean of CFO. Black CFO contours for cumulonimbus (Cb) are also plotted for reference. Starting in the upper-troposphere (UT) we find the largest CFO occur over Africa (AFR), the Indian Ocean, the TWP, and south America (SA). The land masses (AFR and SA) have some of the strongest convective activity, though sporadic, in the tropics [MW92]. The TWP is also marked by frequent strong convection. However, looking at the CFO of tropical Cb, we find maximum CFO of 6% in the boundary layer and free troposphere with CFO values dropping down to 2-3% in the TTL. In the TTL, we find CFO of around 35% which predominately come from tropical thin cirrus. Although the Cb CFO is small compared to the thin cirrus amount, one can see from the large mode of CFO, specifically over the IO and TWP, that the high and Cb cloud contours exhibit similar patterns, suggesting that the cirrus distribution is related to deep convective activity. The large boundary layer CFO close to 30% in the TCP arise primarily from the 2B-CLDCLASS classification of stratocumulus. However, since our data range only captures the deep tropics these clouds are likely trade cumulus misclassified as stratocumulus.
To better distinguish the distribution of CFO in different regions (vertically and horizontally) we subtract out the zonal mean CFO at each pressure level for each longitude bin to get a CFO anomaly cross section (Figure 4.1b). Although discernible in Figure 4.1a, the longitude quadrapole structures observed in cloud amount is more clearly identified in the anomaly field. The regions of free tropospheric positive anomalies over Africa, the TWP, and South America correspond to regions of ascent associated with deep convection. The CFO distribution mimics the relative humidity (RH) structure shown in Figure 3.3c, where the free tropospheric RH follows the water vapor structure while RH in the TTL is mainly governed by temperature. The eastward tilt in the TTL cloud distribution, as well as the RH structure in Figure 3.3c, is a signature Kelvin wave pattern which arises from convective heating (e.g. [VWF10]). To the east of the positive CFO are deficits of CFO in the subsiding branches of the Walker Circulation. In the boundary layer, the ascending regions have negative CFO while the subsidence regions have a preponderance of of boundary layer trade cumulus. This just shows that the subsidence regions of the tropics will typically have a higher CFO of boundary layer clouds than upwelling regions which have a higher CFO of Cb and cirrus.

4.3.1.2 Annual Cycle

Figure 4.2 shows the annual cycle of CFO for all clouds in the 10°S-10°N tropical band. The annual cycle is computed by computing the zonal mean CFO for each month bin over the entire record. Because of the sparse sampling of CloudSat and CALIPSO it is necessary to do a three-month boxcar mean for each month bin. For example, the January and February data bins are comprised of means from (December, January, February) and (January, February,
Figure 4.1: Temporal (Aug, 2006-Dec, 2010) and meridional (10°S-10°N) mean, altitude-longitude (4° longitude bins) cross sections of (a) cloud frequency of occurrence (CFO) in %, CFO contours for cumulonimbus (Cb) clouds (as determined by the 2B-CLDCLASS product) are shown in black in 1.5 % increments; and (b) the anomalies in CFO reported in % as departures from the time record mean CFO at each altitude. Solid pink curves in both panels (near 17 km marks the temporal mean cold-point tropopause height for the same latitude band. Approximate locations of Africa (AFR), the Indian Ocean (IO), tropical western Pacific (TWP), tropical central Pacific (TCP), and South America (SA) are identified.
March) respectively. Figure 4.2a shows the mean annual cycle in CFO for all clouds seen by CloudSat and CALIPSO; the black curve shows the number of cloud profiles available for each month-bin of data. This CFO is the sum of all the CFO of the other cloud types in Figures 4.2b-4.2j, so it represents a CFO relative to every other cloud amount and clear-sky. We do note that the sampling is skewed towards more boreal fall data because 1) the CloudSat and CALIPSO missions didn’t start until mid-boreal summer of 2006, 2) boreal winter data was not available for 2 weeks in 2006 and 3) more than a month of data (between January and February) was missing in boreal winter of 2009-2010. To test whether the reduced sampling impacts our results, we removed data for the entire boreal fall season (SON) season of 2008 and May 2009 to reduce the disparity between the monthly sampling. We then recomputed the annual cycles for each cloud type. This did not make any difference in the CFO distribution. This gives us confidence that the unequal sampling does not change the structure of the annual cycles represented in Figure 4.2.

The total CFO annual cycle in Figure 4.2a shows two main modes of clouds that mainly lie in the UT and the boundary layer with maximum CFO of about 20 %. In the UT the main contributors to the high cloud CFO are the thin cirrus only seen by CALIPSO (Figure 4.2b), though the thicker cirrus and Cb clouds (Figures 4.2c and 4.2d) also contribute. The Cb distribution shows larger CFO during the boreal summer and fall seasons with a reduction in February and March. Although the December and January months have reduced Cb CFO as compared to boreal summer and fall, we do see that the these clouds reach higher into the UT than the Cb in summer and fall, consistent with boreal winter having some of the strongest convective activity (e.g. [NG81]).

The cirrus seen by CALIPSO are persistently higher in altitude than the cirrus
seen by CloudSat by about ∼2 km as shown by the differences in the location of the peak contours for both cirrus classes. This is simply because much of the thick cirrus observed by CloudSat are detrained ice clouds from Cb anvils. The fact that the cirrus seen by CloudSat are mainly detrained from Cb is also indicated by very similar CFO (ranged between ∼1-4 %), in contrast to the roughly factor of ∼5 greater abundance of thin cirrus seen by CALIPSO. The pink curves in Figures 4.2b-d marks the approximate mean cold point tropopause (CPT) height for each month. The seasonality of the CPT is consistent with previous work (e.g. [RG81]).

Although the high clouds (especially the TTL thin cirrus) dominate the tropical CFO, the free tropospheric clouds also contribute significantly in the tropics, though with much lower contribution to the overall CFO (see Figure 4.2a). The mid-level (altocumulus (Ac) (Figure 4.2e) and altostratus (As) (Figure 4.2f)) are mainly mixed phased clouds. Their distributions are roughly uniform but with higher CFO during the boreal summer season. Nimbostratus (Nb) (Figure 4.2g), though they occur relatively infrequently in the tropics, represent the large mesoscale convective systems that contribute a large fraction of the latent and radiative heating, UT moistening, and precipitation [TR10]. Thes systems have distinct peaks in CFO during boreal summer and winter with the winter season having slightly higher CFO.

The boundary-level clouds contribute strongly to the CFO with the a large fraction coming from optically thin clouds seen by CALIPSO only and a smaller fraction coming from the 2B-CLDCLASS classification of cumulus (Cu) (Figure 4.2h) and stratocumulus (Sc) (Figure 4.2j). Given the range of altitudes the Cu clouds reach, it is more likely many of these clouds are congestus versus trade-Cu. The Sc classification is also likely a misclassification since the deep tropics
Figure 4.2: The zonal mean, in the 10°S-10°N latitude band, annual cycle of CFO (abscissa is time in monthly bins), as a function of height (ordinate) for different cloud types, which are identified in each panel. The “All” category represents the sum of CFO from the all the represented cloud types in panels (b)-(j). All cloud types not labeled with “Lidar” are typed by the 2B-CLDCLASS product. The “Cirrus (Lidar)” and other “Other (Lidar)” categories are reserved for clouds identified by CALIPSO alone. Pink curves in panels (b)-(d) are the zonal mean, over the same latitude band as CFO, of cold point tropopause heights (in km). Black curve in (a) shows the total number of cloud profiles available for processing in each month bin. Note that December (D) is the first month in all the panels.
typically do not have a high occurrence of Sc except in the equatorial eastern Pacific [KH93]. These clouds are more likely the shallow trade-Cu which are much more prevalent in the tropics. The “Other Lidar” (Figure 4.2i) category mainly shows evidence of optically thin trade-Cu and mid-level clouds. There may be some thin cirrus in this category with $Z_B$ below 7 km but that will be a very small number and would not contribute greatly to the thin cirrus distribution annual cycle shown in Figure 4.2b.

From these annual cycles, we can compute an estimate of the interannual variability of these cloud regimes. Note: we are able to get a reasonable estimate of the annual cycle because there are enough oscillations in the ENSO cycles to provide contrast between El Niño and La Niña years.

### 4.3.2 Interannual Variability

#### 4.3.2.1 Zonal Mean $\Delta CFO$

To compute the interannual variability of CFO ($\Delta CFO$) we use equation 3.1 in section 3.3.4.1 on the CFO time-record. Figure 4.3a shows the number of counts available for each month bin throughout the record. The CFO time-record (Figure 4.3b), for all clouds observed by CloudSat and CALIPSO, is constructed in the same way as the annual cycle by computing a 3-month running boxcar mean for each time bin. $\Delta CFO$ (Figure 4.3c) for all clouds is then computed by subtracting the annual cycle in Figure 4.2a from the CFO shown in Figure 4.3b.

As with the annual cycle we see that the CPT (pink curve) tracks CFO which just shows the seasonal cycle of both. There are periods when cloud tops do reach into the stratosphere but that occurs relatively infrequently. Looking deeper into the year-to-year variations, we observe interesting cloud patterns in
Figure 4.3: The zonal mean, in the 10°S-10°N latitude band, height-time record of (a) number of cloud profiles, (b) CFO, and (c) ΔCFO for each 1-month bin. Panel (d) are the Ocean Niño Indices (ONI, red curve) and the quasi-biennial oscillation index (QBOI, black curve) with green markers indicating when ONI ≥ |0.5| K (ENSO composites are also denoted in (d)). Pink curves in panels (a)-(c) are the zonal mean, in the same latitude band as CFO, cold point tropopause heights for each 1-month bin reported (in km).
The boundary layer and free tropospheric ΔCFO structure does not seem to correlate with the ONI or the QBOI (Figure 4.3d). One would not expect the QBO to impact the free troposphere since it is mainly a stratospheric process. Furthermore, it is unclear as to whether the ENSO impacts the free tropospheric zonal cloud structure or not. However, from about ∼15 km and above a coherent ΔCFO pattern emerges. There are roughly two biennial cycles of CFO, with negative CFO during boreal winter 2006-2007, 2008-2009, and 2010-2011 and positive CFO during boreal winter of 2008-2009 and 2009-2010. From the beginning of the record to boreal fall of 2007, when the QBOI and ONI are positively correlated, we observed negative ΔCFO in the TTL, particularly near the tropopause (pink curve). When the QBOI and ONI are negative, and in phase, from the fall of 2007 to boreal summer of 2008, we see positive ΔCFO in the TTL. When the QBOI and ONI fall out of phase after the summer of 2008, we see negative ΔCFO for a La Niña with some positive ΔCFO near the tropopause during boreal winter 2009. The strong El Niño of 2009-2010 produced largely positive ΔCFO in the zonal mean in contrast to the relative large El Niño of boreal winter 2006-2007. Boreal winter of 2010-2011 shows strong negative ΔCFO in contrast to the strong La Niña of boreal winter 2007-2008.

We see the main ΔCFO structure are anti-correlated with the QBOI and ONI between the beginning of the time-record to the boreal summer of 2008. After, there seems to be some positive correlation between ΔCFO and the ONI. The CFO changes have a maximum magnitude of ∼4 %, with the largest changes corresponding to latest La Niña (5:E-Q+). This magnitude is mainly due to the changes in thin cirrus (not shown) in the TTL, though the thicker cirrus and Cb contribute if they reach these levels. Looking back at Figure 3.5 we see variations of |∼ 0.5| K, |10%| (or |∼ 0.4| ppmv), and |∼ 10%| in temperature, water vapor, and relative humidity (RH) respectively in the same part of the
atmosphere. Given that the QBO is roughly zonally symmetric, it is possible that the QBO impact on temperature, resulting in $|\sim 10\%| \text{ in RH}$, may have some influence on the observed CFO distribution. To investigate this more, we look at the longitude variations of $\Delta CFO$.

### 4.3.2.2 Longitudinal Variation of $\Delta CFO$

Figure 4.4a and 4.4b show the CFO and $\Delta CFO$ at 16.80 km for all high cloud amount (clouds shown in Figure 4.2b-d). The CFO distribution itself shows the migration of cloud between the TWP and TCP (near the dateline) corresponding to the oscillating phase of ENSO. There are periods when $\Delta CFO$ becomes roughly zonally symmetric. Looking at the longitudinal structure between the black parallel bars in Figure 4.4b we see approximately zonally symmetric negative (label Q1), negative (label Q2), positive (label Q3), and negative (label Q4) anomalies; the ENSO event of 1:E+Q+ (label Q1) does have positive $\Delta CFO$ in the TCP since the TTL temperatures are colder during El Niño. However, the rest of the tropics during this period has negative $\Delta CFO$, consistent with the westerly (warm) phase of the QBO. Q2 shows a very coherent band of negative $\Delta CFO$ during at time when the ONI is zero and the QBOI is at a westerly maximum. Q3 shows positive $\Delta CFO$ consistent with a easterly (cold) phase of the QBO. Q4 shows negative $\Delta CFO$ across the entire tropics, consistent with the westerly QBO. It should be noted that because the record is relatively short, it is difficult to quantify how much the signals are related to the QBO or ENSO. However, the observed $\Delta CFO$ patterns are consistent with a QBO imposed temperature structure (seen in Figure 3.5a) on the tropopause region, i.e. a westerly (easterly) QBO phase leading to warm (cold) temperature anomalies ($\Delta T$), leading to reduced (increased) $\Delta CFO$. 

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Figure 4.4: Longitude Hovmöller diagrams (4° longitude bins) of (a) CFO and (b) ΔCFO for the entire tropical band between 10°S-10°N at 16.32 km. The same tropical locations as in Figure 4.1 are referenced. Panel (c) shows the ONI and QBOI for reference with the four ENSO composites summarized in Section 4.2.3. Dashed-black contours in (b) mark zero CFO. Q1-Q4 labels indicate periods when ΔCFO are roughly zonally symmetric.
In Figure, 4.4b we also see the zonally asymmetric impacts of ENSO. These are easily discernible and follow the conventional view of the migration of clouds between the TCP and TWP resulting from the changes in the Walker Circulation. The TCP (TWP) is colder (warmer) during El Niño events so we expect positive (negative) $\Delta CFO$ in the TCP (TWP) and the reverse behavior during La Niña cycles. Thus, the obvious ENSO signature is a zonally asymmetric change on cloud amount mainly confined to the tropical pacific. There is a peculiar difference between 2:E-Q- and 5:E-Q+. Though both La Niñas have very similar anomalies between -1.4 to -0.6 K (for 9 and 8 months for 2:E-Q- and 5:E-Q+ respectively), $\Delta CFO$ is positive for 2:E-Q- and negative for 5:E-Q+ in the TWP and the zonal mean. In Section 3.3.4.3 we found that the QBO and ENSO served to enhance (reduce) temperature anomalies ($\Delta T$) in the TWP (TCP) when these modes were in phase and vice-versa, i.e. enhanced (reduced) $\Delta T$ in the TCP (TWP), when the QBO and ENSO are out of phase. Event 2:E-Q- corresponds to a La Niña in phase with a QBO easterly which corresponds to the TWP having enhanced cold anomalies, which is consistent with the positive $\Delta CFO$ in the TWP. However, because the QBO and ENSO are out of phase for 5:E-Q+, we see enhanced negative $\Delta CFO$ in the TCP and reduced anomalies in the TWP (now negative). Although this changing behavior is only suggestive, given that we only have two events that clearly show this shift in $\Delta CFO$, it is consistent with the overall thermodynamical structure expected from the interaction between the QBO and ENSO.

To capture the vertical and horizontal structure of cloud amount, for all clouds, we compute height-longitude cross-sectional composites of $\Delta CFO$ over the ENSO events described in Section 4.2.3. Figure 4.5a-e show much more clear dipole structure in $\Delta CFO$ where almost all clouds in the tropics respond to ENSO. From 0 to $\sim$15 km $\Delta CFO$ is roughly vertical. Above $\sim$15 km the cloud
structure tilts vertically eastward, pointing to a Kelvin wave signature in the high cloud structure on interannual time-scales (e.g. [VWF10]). This is consistent with Rossby and Kelvin circulations induced by tropical equatorial heating (e.g. [Gil80, HH98]). Although this mainly impacts the thin cirrus $\Delta CFO$, the thick cirrus clouds are also impacted at altitudes above 15 km (not shown). The temperature patterns in in Figures 3.11c and 3.11d both show eastward tilting temperature anomalies resulting from the convective heating induced by ENSO. Although changes in $H_2O$ amount impacts RH in the TTL, the larger impacts on RH come from temperature (see Figure 3.3c).

From these vertical cross sections, one can qualitatively identify the strength of the ENSO event. For example, differences are apparent for the two La Niña events. 2:E-Q-, consistent with the larger ONI, show clouds reaching far westward into Africa while the anomalies for 3:E-Q+ only reach the Indian Ocean. Both do show increased $\Delta CFO$ in the eastern sector over South America, whereas both El Niño events show negative free-tropospheric and boundary $\Delta CFO$ over the same region. The most recent El Niño shows very strong high cloud $\Delta CFO$, roughly a factor of 2 greater $\Delta CFO$ than any other ENSO event in this time-record. We see this is intimately connected with the strength of convection as evidenced by 4:E+Q* having a factor of 2 greater rain rate anomalies (black curves on right ordinate in each panel) than the other ENSO events.

Looking more closely at the boundary layer we see $\Delta CFO$ dipole patterns that are opposite to the free-troposphere and TTL, similar to the patterns seen in Figure 4.1b. During El Niño the TCP (TWP) has negative (positive) $\Delta CFO$ in the boundary layer while above there are positive (negative) anomalies over the same region. This shows that when TCP (TWP) convective active increases (decreases), the boundary layer cloud amount decreases (increases) since the os-
Figure 4.5: Height-longitude cross sections (4° longitude bins) of boreal winter (DJF), interannual mean composites of $\Delta CFO$ for the following events (a) 1:E+Q+, (b) 2:E-Q-, (c) 3:E-Q+, (d) 4:E+Q*, and (e) 5:E-Q+. Latitude band taken between 10°S-10°N. Pink curves show composite mean of CPT for the same period and latitude band in each panel. Black curves (right ordinate) shows the interannual rain rates derived from TRMM. The same tropical locations as in Figure 4.1 are referenced.
cillation of ENSO modulates an oscillation in cloud regime.

The coherency in the ENSO asymmetric impact on clouds is clearly observed. The entire cloud distribution, including all mid-level clouds are also modulated by tropical sea surface temperature (SST) changes. However, the coherence of the ENSO cloud impact may be more a result of how the mid-level clouds are likely connected to convective activity and the low level clouds are associated with changes in subsidence linked to the changes in location of convective activity. One thing to note in the 5:E-Q+ composite is the coherent negative $\Delta CFO$ strip centered around the mean tropopause (pink curve). This, as previously discussed, suggests a QBO impact on the $\Delta CFO$ structure. While we see in 2:E-Q- eastward tilting positive $\Delta CFO$ tilting and penetrating the stratosphere, the eastward tilting positive $\Delta CFO$ in 5:E-Q+ do not even reach the tropopause. This may suggest a QBO impact on the Kelvin waves that produce the tilting $\Delta CFO$ patterns. More analysis is required to substantiate this claim.

4.3.3 Statistical Significance

We have quantified the ENSO asymmetrical impact on $\Delta CFO$. Furthermore, we have found that almost all clouds in the tropics are impacted by this. We’ve also found coherent structures in the zonal mean $\Delta CFO$ that seem to follow the QBOI and ONI depending their relative phase to each other (Section 4.3.2.1). To further explore this, we compute statistical correlations between the zonal mean $\Delta CFO$ and QBOI and ONI. Since the indices and $\Delta CFO$ are independent parameters, we do not need to worry about one variable influencing the other statistically.

Figures 4.6a-d show the correlations for various cloud regimes for the period that spans between Aug, 2006 - July, 2008 (24 months) or almost one full QBO
cycle when the ENSO and QBO are in phase (hence the label E24 and Q24 for the ENSO and QBO correlations respectively). For all high clouds (Figure 4.6a) both Q24 and E24 show high anti-correlations with maximum correlation coefficients of $R_{Q24} = -0.91$ and $R_{E24} = -0.81$ (at 16.32 km). They are statistically significant to the 99.99% level (as indicated by the gray (Q24) and pink (E24) patched regions). This behavior is also seen in Figure 4.6c ($R_{Q24} = -0.84$ and $R_{E24} = -0.71$ for cirrus seen by CloudSat alone at 16.56 km) and Figure 4.6d ($R_{Q24} = -0.90$ and $R_{E24} = -0.82$ for cirrus seen by CALIPSO alone at 16.32 km). A reduced correlation of $R_{Q24} = -0.74$ (at 16.8 km) is computed for Cb clouds seen in Figure 4.6b. However, this anti-correlation is consistent with the results of [CHM98, CMH02] who found evidence that the westerly (easterly) QBO phase stabilized (destabilized) the TTL, thus decreasing (increasing) the vertical reach of deep convective clouds. The correlation with the ONI shows some anti-correlation ($R_{E24} = 0.51$ but this is not statistically significant at the 99.99% confidence level.

Now including the rest of the time series, where each calculation must have at least 50 months of data, we immediately find that the only statistically significant correlations with $\Delta CFO$ are with the QBOI, though with reduced correlation coefficients of $R_{Q50} = -0.62$ (at 16.32 km and 16.5 km), -0.46 (at 16.56 km), -0.56 (at 16.32 km) and -0.61 (at 16.08 and 16.32 km) for all clouds, Cb, thick cirrus (CloudSat), and thin cirrus (CALIPSO) respectively. Roughly the same correlation coefficients result from computing the correlations as function of distance below the zonal mean CPT height for each month in the time-record shown in Figure 4.3, though with reduced correlation with $R_{cpt50} = -0.54$ (All Clouds), -0.42 (Cb), -0.51 (CloudSat cirrus), and -0.53 (CALIPSO cirrus) at 1200 m, 240 m, 240 m, and 1200 m below the CPT respectively. The only statistically significant correlations are for the “All Clouds” case and thin cirrus (CALIPSO), which is consistent with temperature variances impacting the thinnest of clouds in the
Figure 4.6: Correlation coefficients of QBOI (black curves marked Q24 and Q50) and ONI (red curves marked E24 and E50) with CFO as a function of height between $\sim$15-18 km (in 240 meter increments). Panels (a) - (d) show coefficients for the time period when the QBO and ENSO are in phase (roughly between Aug, 2006-Jun, 2008); these are marked Q24 and E24. Panels (e)-(h) are coefficients for the entire time-record marked Q50 and E50. Panels (i)-(l) are coefficients for the entire time-record as a function of distance below the CPT. Colored patches denote the range of altitudes for which the correlation between $\Delta CFO$ and both the QBOI (gray) and ONI (pink) are statistically significant at the 99.99 % level or better. Dark pink patches correspond to heights when both correlations are statistically significant.
tropopause region. The reduced correlation may be a result of not sampling, i.e. co-locating in space and time, the A-Train data with the reanalyses pixel by pixel; the CPT from the reanalyses may capture more processes (since it has sampling at every 0.25 × 0.5 degree grid box every 6 hours) than the A-Train data, which is aliased to ∼13:30 (ascending orbit) and ∼01:30 (descending orbit).

The cause for the reduced correlation beyond the first cycle of the QBO (E24 and Q24) can be qualitatively identified in Figure 4.3. When the QBO and ENSO are in phase for the first 24 months of the record, ∆CFO follows the QBOI very closely. Beyond, there is a mixture of positive correlation between +ONI with +∆CFO and -QBOI with +∆CFO between boreal summer of 2009 and boreal winter of 2011.

Although the correlations do suggest a QBO influence on tropical high cloud amount, we cannot conclusively show they are related because the time record is simply too short (only two full QBO cycles). Furthermore, we cannot conclude that the ENSO does not have a zonally symmetric influence on cloud amount from the lack of statistical significance in Figures 4.6e-l, for the same lack of a longer time series. However, what we have demonstrated statistically with this short record is that the QBO can serve as one possible explanation for the zonally symmetric coherent patterns in high cloud amount near the tropopause.

4.4 Discussion

We have quantified the time-record mean and annual cycles of CFO. From both calculations we found CloudSat and CALIPSO capture well-known structures of the tropical cloud distributions. This provided confidence in understanding the seasonal distributions of clouds which are essential for de-seasonalizing the CFO
structure, i.e. quantify the interannual variability of CFO ($\Delta CFO$). From the longitudinal varying and zonal symmetric structure of $\Delta CFO$, we were able to conclude that: the ENSO impacts the entire tropical cloud distribution with high thin cirrus experiencing the largest variation, and the QBO is a likely contributor to the tropical zonal mean $\Delta CFO$. The statistically significant (at the 99.99% confidence interval) correlation between $\Delta CFO$ and the QBOI both the 24 and 50 month time-record further bolsters this result. A longer time series with many more QBO and ENSO cycle would greatly aid in determining to what degree both modes contribute to the TTL cloud structure.

From this study, we found that thin cirrus, i.e. cirrus observed only by CALIPSO, are the greatest contributors to the CFO and $\Delta CFO$ distribution in the deep tropics. This points to possible QBO impacts on the tropical radiative balance. It has been shown that high cloud amount can have a significant impact on the level of zero radiative heating (LZRH) [CLP05] which marks the approximate tropical altitude where vertical transport transitions from being mainly convective to radiatively driven. Though, the zonal mean $\Delta CFO$ magnitude only reaches $\sim 4\%$, the zonal region-to-region magnitudes greater than $\sim 15\%$ in the TTL. What may help in quantifying the radiative impacts of these variations of $\Delta CFO$ is to derive radiative heating profiles for each of the CloudSat and CALIPSO observations. Currently, the CloudSat team provides a 2B-FLXHR (flux and heating rate) product for clouds observed by CloudSat only. Thus, it will be useful to derive similar heating rate profiles including observations from CALIPSO since these clouds are the main contributors to the tropical CFO and $\Delta CFO$ distribution.
CHAPTER 5

Radiative Heating of Tropical High Clouds

“Happiness is not a matter of intensity but of balance, order, rhythm and harmony.” – Thomas Merton

5.1 Preamble

The radiative balance of the tropical atmosphere is largely determined by the balance of longwave cooling of free and upper tropospheric water vapor (H$_2$O) and infrared absorption (heating) by ozone (O$_3$) and carbon dioxide (CO$_2$) in the tropical tropopause layer (TTL, between 70-150 hPa) [TC02]. Outside deep convection, a parcel must be lifted to this level of zero radiative heating (LZRH, $Q_{net}$ = 0) in order for the background atmospheric net radiative heating to lift it into the stratosphere [GFM04]. When parcels do reach the tropical tropopause layer (TTL), which is predominantly above the LZRH, residence times of parcels reach a few weeks [FWP04]. Previous estimates of the LZRH put this level around ~15 km [FOT00, GFM04, CLP05] or ~135 hPa [She00]. The “All-Sky” LZRH, i.e. when one includes the heating/cooling contribution from clouds, drops by ~0.5-1 km [GFM04, CLP05]. These estimates were either derived from model and/or a sparse network of in-situ profilers (e.g. radiosondes and balloon sonde). In this work, we use an approach similar to [GFM04] and [CLP05] by using a ra-
diative transfer model to compute heating rate profiles. However, we use global soundings of H$_2$O, temperature, and O$_3$ from AIRS and MLS (H$_2$O only) to better quantify the global structure of the LZRH for clear-sky (LZRH$_{clear}$) and “All-Sky” (LZRH$_{all}$).

5.2 Data and Methodology

5.2.1 A-Train Soundings and Cloud Profiles

In order to calculate the radiative fluxes via a radiative transfer model (RTM) information about the atmospheric and cloud state are required. Our data spans the 40$^\circ$S-40$^\circ$N latitude band, however our analysis will primarily focus on the 10$^\circ$S-10$^\circ$N tropical band. At the time of this work the only period for which AIRS, MLS, Cloudsat, and CALIPSO were co-located was between May 2008-February 2009. Atmospheric temperature, O$_3$ and pressure are provided by the Atmospheric Infrared Sounder (AIRS). H$_2$O come from the merged profiles described in Chapter 2. In addition, in order to put the cloud profiles on a pressure grid, we use the AIRS geopotential height profiles to translate the cloud height dimension to pressure. We only compute data over ocean; this is determined from the AIRS land fraction parameter ($\sim$15 km resolution). We limit our analysis to the ocean because the land surface albedo properties are highly variable and not well constrained. The total number of cloud profiles used for the RTM calculations between 40$^\circ$S-40$^\circ$N is 3,873,476; the tropical total (10$^\circ$S-10$^\circ$N) is 1,160,582.

Cloud profiles come from an older version (version 2) of the CloudSat 2B-GEOPROF-LIDAR [MZV09] described in Chapter 4 (version 3) (the work done in this chapter was actually done before much of the work in Chapter 4). We ex-
clude all pixels with mix phased clouds: altostratus, altocumulus, nimbostratus, and deep convective clouds as classified by the Cloudsat 2B-CLDCLASS [SW08] product to avoid the ambiguity and difficulty of partitioning between liquid and ice hydrometeors in these clouds. Thus, even for scenes with high level ice clouds or shallow water clouds, if there are any mix phased clouds, we exclude those pixels. Furthermore, clear-sky is defined as pixels for which AIRS, CloudSat, and CALIPSO indicate there are no clouds. An “All-Sky” scene for this study will only include clear-sky and cloudy scenes, as seen by CloudSat and CALIPSO, with no mid-level, mix phased clouds. We acknowledge that this may bias the “All-Sky” results by removing effects from thicker mix phase clouds. However, the goal of this study is primarily quantify the separate heating effects of the cirrus independent of other clouds. As in Section 4.2.2, further quality control and classification was necessary to separate the clouds observed by Cloudsat from those profiled by CALIPSO. It should be noted that because the cloud profiles are selected for pixels for which AIRS and MLS have successful soundings, these scenes primarily contain clouds of reduced cloud fraction (see Figure 3.1 for AIRS and MLS sampling statistics).

Furthermore, in order to achieve the proper behavior in the radiative heating profiles we divide up the atmospheric and cloud profiles such that every cloud layer has at least three pressure grid levels, one for the top, middle, and bottom of the cloud. For example, without doing this procedure, thin cirrus clouds will not exhibit the proper heating and cooling at their bottoms and tops respectively. This procedure was done successfully by using a hybrid pressure grid that includes all pressure levels for both the AIRS and MLS grids. This leads to pressure layers that have a maximum vertical separation of $\sim 14$ hPa in the boundary layer and $\sim 4$ hPa at 100 hPa. This corresponds to a geopotential height layer difference of a maximum of $\sim 200$ meters at 100 hPa and $\sim 100$ meters in the
boundary layer. After, the cloud profiles derived on the native Cloudsat altitude grids are placed on the new pressure grid derived from the previous sub-dividing procedure. Note: This version of the 2B-GEOPROF-LIDAR, specifically due to the CALIPSO product, had a global bias in subtropical boundary layer cumulus (single layer clouds) cloud fraction due to misclassification of aerosol layers as clouds [VKT10]. This, however, will not significantly impact our conclusions as our main results primarily focus on the impact of high clouds on the tropical radiative balance within the 10°S-10°N band.

5.2.2 Radiative Transfer

We use the Fu-Liou correlated-K broadband radiative transfer model (RTM) described in [FL92, FL93] to compute heating rate profiles. Concentrations of gases CO₂, CH₄, and N₂O are 380 ppmv, 1.775 ppmv, and 0.32 ppmv respectively. The model parameterizes other gases as well: CO (0.16), O₂ (2.09×10⁵), NO (0.0005), SO₂ (0.001), NO₂ (0.001), and CH₃Cl (0.5×10⁻³), where the parenthetical values are the concentrations in units of ppmv. We compute the surface albedo using the ocean surface albedo parameterization described in [JCS04]. This model computes a four-dimensional surface albedo look-up-table (LUT) based on chlorophyll concentration, ocean surface wind speed, aerosol optical depth/cloud optical depth, and solar zenith angle. We assume a chlorophyll concentration of 0.2 mg/m³ which is about the global average [JCS04]. We use a constant ocean wind speed of 9 m/s based on the global distribution of QuikSCAT measured ocean wind speeds shown in [CS08] (this value is the mode of the distribution). We assign a LUT optical depth (cloud and aerosol) of 0.0 because we want to model albedo to mimic the albedo before the RTM propagates the downward and upward impact of clouds, aerosols, and the atmosphere. Solar zenith angle (SZA)
is obtained from the AIRS data. The albedo LUT value is then linearly interpolated, in the SZA dimension, to the AIRS SZA. We do not include any specific aerosol information other than assume a background aerosol optical depth of 0.2.

Finally, a vertical distribution of cloud microphysical properties is necessary, i.e. ice water content (IWC), liquid water content (LWC), and particle size ($D_e$). For this work, we assume a constant extinction in the cloud layer and a 100% cloud fraction within the CloudSat/CALIPSO footprint. This cloud fraction assumption is reasonable given the small 1 km footprint of the CloudSat instrument (the GEOPROF-LIDAR product is reported on the CloudSat horizontal grid). At the time of this work, the microphysical retrievals derived from the joint Cloudsat and CALIPSO cloud profiles have not been validated. Thus, in order to calculate radiative fluxes and heating rates, we parameterize the cloud water contents based on the recipe shown in Table 5.1. The ice cloud values are based on in-situ measurements taken from [HP84]. A water droplet radius of 10 µm is assumed based on [Har02]. $D_e$ is computed using the following equation from [LGY08]:

$$\ln(D_e) = a + b \ln(IWC) + c(\ln(IWC))^2,$$  

(5.1)

where (a, b, c) is (5.4763, 0.55175, 0.02693). The resulting $D_e$ are also listed in Table 5.1. Although [LGY08] compute coefficients (a, b, c) for the mid-latitude as well, we use the same aforementioned coefficients for the entire 40°S-40°N latitude band. The exact values of IWC and LWC, so long as they are reasonable estimates for the clouds of interest, is not crucial for this study as the motivation of this study is to test which ice clouds have the largest impact on the tropical radiative balance given reasonable IWC and LWC values.
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<th>$D_e$ ($\mu$m)</th>
<th>T Range (K)</th>
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<td>$288 &lt; T &lt; 298$</td>
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</tbody>
</table>

Table 5.1: Table of parameterized liquid water (LWC), ice water content (IWC) [HP84], and particle size ($D_e$) [LGY08] as a function of temperature. Warm cloud particle size is 20 $\mu$m [Har02]. LWC = 0.3 g/m$^3$ taken from Fu-Liou RTM. Note: no mixed phase clouds are parameterized.
5.3 Results

5.3.1 Humidity and Radiative Heating Distributions

In this section, the meridional structure of the zonal mean relative humidity (RH) and net radiative heating \( Q_{\text{net}} = \text{shortwave heating} (Q_{\text{SW}}) + \text{longwave heating} (Q_{\text{LW}}) \) are shown between the latitude belt of 40°S-40°N. The RH is computed using the recipe described in Section 3.2.1. Figure 5.1 shows this meridional distribution for a) clear-sky, b) CloudSat only scenes without any CALIPSO identified clouds, and c) CALIPSO scenes without any Cloudsat profiles clouds. One observation is that cloudy scenes are more moist than the clear-sky scenes with the CloudSat scene (Figure 5.1b) having a strong mode of high humidity throughout the upper troposphere (UT); a thinner moist layer centered around 150 hPa also exists for the CALIPSO scenes as well (Figure 5.1c). However, the conditions in the free troposphere are also more moist for both cloudy (CloudSat and CALIPSO) scenes. The boundary layer RH is quite constant for all scenes, particularly in the tropics. For all scenes, there is a strong band of high RH centered around 100 hPa (approximately around the tropical tropopause) within the latitude belt of 20°S - 20°N. This is primarily due to the strong impact of temperature on the saturation mixing ratio at these pressure levels. For the cloudy scenes, the RH distribution qualitatively follows the cloud frequency of occurrence (CFO) distribution (dotted contours in Figure 5.1b and 5.1c). For the CALIPSO maps the RH mode around the tropopause is also coincident with non-negligible CFO of optically thin cirrus clouds. As will be shown, these clouds contribute significantly to the radiative balance of the tropical UT.

Figure 5.2 shows a similar meridional distribution but for \( Q_{\text{net}} \). Figures 5.2a, 5.2b, 5.2c, and 5.2d show \( Q_{\text{net}} \) for clear-sky, CloudSat only scenes, CALIPSO
Figure 5.1: Meridional distribution of the time record zonal mean RH for: a) clear-sky, b) CloudSat only scenes, and c) CALIPSO only scenes. Latitude and Pressure are on the abscissa and ordinate respectively. Dotted contours show cloud frequency of occurrence (CFO) for CloudSat (panel b) and CALIPSO (panel c) observed clouds. See text for description of scene construction.
Figure 5.2: Meridional distribution of the time record zonal mean net (short-wave+longwave) heating for: a) clear-sky, b) CloudSat only scenes, c) CALIPSO only scenes, and d) All-Sky. Latitude and pressure are on the abscissa and ordinate respectively. CFO (dashed contours) are plotted for CloudSat (panel b), CALIPSO (panel c) and “All-Sky” (panel d)See text for description of scene construction.
The dark curves in Figure 5.2a can be interpreted as the meridional distribution of LZRH\textsubscript{clear}. Within the tropics, and even the extra-tropics, LZRH\textsubscript{clear} is quite uniform approximately at an isobar of $\sim$140 hPa. Below this level the dominant source of negative heating rates arises from the longwave cooling from H$_2$O. As H$_2$O amounts fall off quickly in the TTL the small longwave heating contributions by O$_3$ and CO$_2$ (and perhaps other greenhouse gases) result in a net warming.

Figures 5.2b and 5.2c show this same meridional distribution for the Cloud-Sat (5.2b) and (5.2c) CALIPSO cloud regimes, i.e thick and thin cirrus regimes respectively. At first glance it is obvious the thicker cirrus clouds, with IWC concentrations of at least an order of magnitude greater than the clouds (as shown in Table 5.1), contribute much higher magnitudes of $Q_{net}$ than the optically thinner clouds observed by CALIPSO. The maximum heating takes place slightly below the 200 hPa pressure level, the approximate location of neutral buoyancy in the tropical atmosphere. This is not surprising as much of the observed thick cirrus clouds correspond to detrained anvils from deep convective clouds. The location of peak heating for the thinner clouds (observed by CALIPSO) is $\sim$125 hPa. The thicker clouds, in the climatological zonal mean, heat the atmosphere about a factor of 2-3 more than the thin clouds do. Overlaid on the $Q_{net}$ composites are contours of CFO. The $Q_{net}$ and CFO distribution correlate well, indicating the clouds are indeed a significant source of heating at these levels. As will be shown later, the vertical distribution of clouds is important in determining the location of LZRH\textsubscript{all}.

The “All-Sky” LZRH\textsubscript{all} composites show a meridionally varying LZRH structure approximately following the CFO distribution. The northern hemisphere (NH) shows a LZRH\textsubscript{all} at a lower altitude than in the southern hemisphere (SH).
At first, this may seem surprising as the tropical western Pacific (TWP), in
the SH, should have a substantial impact on LZRH\textsubscript{all} due to the large coverage
of thin cirrus in this region due to this region being the coldest in the tropics
[CAT04, LEG11]. This asymmetry, however, is likely due to the data record
missing data in March and April when TWP temperatures in the TTL are com-
mensurate with the TWP temperatures in boreal winter. Boreal summer (June,
July, August) convection is strongest in the Indian Monsoon region and thus can
explain the particularly strong heating that results in the dip in LZRH\textsubscript{all} in the
NH.

5.3.2 Humidity and Heating Anomalies

Previously we quantified the general RH and LZRH\textsubscript{clear}/LZRHA\textsubscript{all} structure and
identified differences between clear and cloudy sky conditions. To further explore
these differences we subtract the composite shown in Figure 5.2d from those in
Figures 5.2a, 5.2b, and 5.2c. This procedure produces $Q_{\text{net}}$ anomalies relative to
“All-Sky” conditions ($\delta Q_{\text{net}}$, Figures 5.3a-c). The identical differencing is also
done on the temperature ($\delta T_{\text{net}}$, Figures 5.3d-f) and specific humidity ($\delta q_{\text{net}}$,
Figures 5.3g-i) fields. All figures also show the contours of CFO for reference.

For $\delta Q_{\text{net}}$ it is apparent that the cloud scenes have much stronger warming
than the clear-sky conditions. As in Figure 5.2 the heating of the thick cirrus
(CloudSat) are at a lower altitude than the thinner clouds (CALIPSO). However,
the anomalies show that the thicker clouds produce more heating throughout
the entire free troposphere compared to the clear-sky and thin cloud conditions.
This is due to a more moist atmosphere associated with tropical deep convection
[LEG11]. This suggests that the conditions for which only thin TTL cirrus ex-
ists are consistently drier than conditions for which thick cirrus form. We note
Figure 5.3: Meridional distribution of time record zonal mean anomalies of: (a-c) net radiative heating, (d-f) temperature, and (g-i) water vapor mixing ratio (in %) for clear-sky (column 1), Cloudsat only scenes (column 2), and CALIPSO only scenes (column 3). Anomalies are defined as the differences between each of the clear or cloudy and “All-Sky” composites. Latitude and pressure are on the abscissa and ordinate respectively. CFO contours (dashed) are also plotted for CloudSat (panels b,e,h) and CALIPSO (panels c,f,i). See text for description of “All-Sky” composite construction.
that there are scenes that can have both thick and thin cirrus (these are in the “All-Sky” composites), but we have removed these cases from the CloudSat and CALIPSO only scenes to narrow down the impact of the clouds identified by these instruments.

The $\delta T_{net}$ distribution shows peculiar behavior that requires some explanation, particularly in the TTL. The clear-sky $\delta T_{net}$ shows higher tropical anomalies than for either of the cloudy composites. Although the cloudy scenes show heating in the regions of high CFO, $\delta T_{net}$ is negative. Also, within the cloudy scenes (Figures 5.3e and 5.3f) the free troposphere is warmer than in the clear-sky composites. The exact mechanism for why negative $\delta T_{net}$ are co-located in these cloudy regions above a warm free troposphere is still a topic of research [HG01, HHF01, SD03, HN07]. Figures 5.3c and 5.3f are consistent with cold temperatures leading to more cloud amount (as indicated by the CFO contours), and the resulting clouds lead to radiative heating in the TTL which leads to ascent of air parcels. The typical vertical velocities in the TTL are $\sim0.5$ mm which indicates that to transport a parcel from LZRH$^\text{clear}$ ($\sim15$ km) to the tropical tropopause (at $\sim18$ km) it would take about 2 months in clear conditions. The addition of these thin clouds can significantly enhance the vertical velocities, due to the heating by clouds, and reduces the transit time to about two weeks [CLF06].

Finally, we investigate the $\delta q_{net}$ structure. In both cloud cases (Figures 5.3h and 5.3i), the free troposphere and lower portion of the UT (up to $\sim150$ hPa) are more moist than in clear-sky. This suggests that while we have chosen only scenes without deep convection (due to the AIRS and MLS sounding limitations) the effects of convection are still evident. Previously we found that $\delta T_{net}$ (Figures 5.3e and 5.3f) showed warm anomalies in the free troposphere, consistent with convective heating. [LEG11] also showed that on interannual time-scales the
convective heating is robust in regions where active convection (with associated moist anomalies) takes place, even though convection is not explicitly sampled in the temperature composites (see Figures 3.3 and 3.11). Above these moist regions, particularly in the CALIPSO composite, are areas of dry $\delta q_{net}$ coincident with the observed clouds; see region just south of the equator between latitude $10^\circ$S-0. The clear-sky composite, on the other hand, shows moist anomalies in the same region. This indicates a possible signature of dehydration of TTL H$_2$O due to horizontal advection through cold regions where TTL cirrus are ubiquitous [HG01, CAT04].

5.3.3 Tropical Radiative Balance

In the previous sections we quantified the meridional structure of net heating as well as the differences between clear-sky and cloud regions from a radiative and thermodynamical perspective. However, we did not explicitly quantify the differences in the net heating due to the different (CloudSat and CALIPSO) cloud regimes relevant to this work. In this section we place our focus on the tropical belt ($10^\circ$S-$10^\circ$N) and composite all RH and $Q_{net}$ profiles over the entire region to quantify, in the mean, the contribution of different clouds to the $Q_{net}$ distribution and, thus, LZRH$_{alt}$.

Figure 5.4a shows the complete tropical mean RH profile partitioned by clear-sky, cloudy-sky, and “All-Sky” conditions (see legend Figure 5.4). At all pressures, any composite with clouds (CloudSat, CALIPSO, and, by default, “All-Sky”) have greater RH than clear-sky. In the free troposphere, and even in the lower parts of the UT (around ~200 hPa), this indicates that cloud conditions are associated with a environment with more specific humidity. In the TTL, cloudy conditions are associated with cold temperatures, resulting in lower saturation
mixing ratios which also raise RH. Another interesting feature is the location, \( \sim 160 \) hPa, where RH (CloudSat) and RH (CALIPSO) intersect. Below this pressure level the CloudSat RH noticeably dominates, presumably due to higher specific humidity concentrations in scene with thicker cirrus clouds; remember that these thick cirrus clouds are predominantly detrained anvil cirrus from tropical deep convective clouds. Above 160 hPa the strong temperature effect on the saturation mixing ratio increases the RH (CALIPSO) above RH (CloudSat).

In the same figure, the mean tropical CFO (gray dashed curve) is shown for reference.

In Figure 5.4b the \( Q_{net} \) composites are plotted. Also included are the LW contributions from the CloudSat and CALIPSO scenes which qualitatively mimics the corresponding RH profiles. The clear-sky LZRH\(_{clear} \) (solid black curve) is \( \sim 140 \) hPa, with a max \( Q_{net} \) of about \( \sim 0.4 \) K/day around the tropopause, presumably due to the longwave heating of CO\(_2\) and O\(_3\). Our LZRH\(_{clear} \) is about 20 hPa lower in the atmosphere than the mean values in [GFM04]. However, in [GFM04] the LZRH\(_{clear} \) was \( \sim 140 \) hPa for noon-time simulations. Our results are consistent with their noon results since the A-Train ascending equatorial crossing time is 13:30. This implies a diurnal cycle in LZRH\(_{clear} \).

To quantify the impacts of clouds we compute the “All-Sky” (all profiles) composites and subtract out the CALIPSO (blue “No Lidar” and “CloudSat LW” curves in Figure 5.4b) or CloudSat (red “No Radar” and “CALIPSO” LW curves in Figure 5.4b) clouds to show their contributions to \( Q_{net} \) and shifting LZRH\(_{clear} \). For the case of removing CALIPSO clouds (“All-Sky” (No Lidar) blue curve) we find the maximum \( Q_{net} \) does not change very much around the tropopause. This is due to the negligible longwave contribution observed on CloudSat only scenes. In the UT (between \( \sim 175-300 \) hPa, the longwave contribution from CloudSat ob-
Figure 5.4: Time record mean of (a) RH and cloud frequency of occurrence, and (b) radiative heating for all profiles in the tropical belt between 10°S-10°N. Legend specifies particular composites. Cloudsat LW and CAIPSO LW show just the longwave contribution the cloud heating.
servations is quite large due to the strong heating (max longwave heating of $\sim 2.5$ K/day) of detrained thick cirrus clouds in this layer. This heating contribution does shift $Q_{net}$ to the right enough to move the $\text{LZRH}_{\text{clear}}$ down to $\sim 150$ hPa. Because of the location of these clouds, the change in the $\text{LZRH}_{\text{clear}}$ is a modest $\sim 10$ hPa.

Subtracting only the CloudSat clouds (“All-sky” (No Radar) red curve) has a more dramatic impact on $\text{LZRH}_{\text{clear}}$. Due to the ubiquity and vertical location of the clouds observed by CALIPSO, the maximum $Q_{net}$ shifts to the right about $\sim 0.2$ K/day leading to in $\text{LZRH}_{\text{clear}}$ shifting down to $\sim 170$ hPa, which corresponds to a pressure change of 30 hPa from the clear-sky case. A second local $Q_{net}$ peak arises at $\sim 125$ hPa due to the longwave heating of the optically thin cirrus. Once again, it is the longwave heating, now in the TTL, that shifts $\text{LZRH}_{\text{clear}}$ so far down in the atmosphere. When including all profiles (green “All-Sky” (Total) curve), $\text{LZRH}_{\text{clear}}$ moves down to $\text{LZRH}_{\text{all}} = 180$ hPa, 40 hPa or $\sim 1.5$ km, down from $\text{LZRH}_{\text{clear}}$. This lowering (in altitude) of $\text{LZRH}_{\text{clear}}$ is about 0.5 km greater than the estimates presented in [CLP05]. This large difference is likely due to the difference in the cloud inputs into the RTM. While [CLP05] had estimates of cloud top pressure and optical depth from International Satellite Cloud Climatology Project (ISCCP) [RS99], they assumed all clouds were a single layer assigned with a climatological distribution of vertical extent (e.g [Lio92]). Thus, each cloud layer is assigned an optical depth for the whole layer not distinguishing between multi-layer and single layer systems, whereas the CloudSat and CALIPSO data provide the vertical extent of each individual cloud within a layer. Since the atmospheric column optical depth (minus the influence of aerosols) is predominately from clouds, assuming a single optical depth for an entire layer of a multi-cloud system may, in effect, reduce the optical depth of each cloud layer. This is important as we have shown that not only does the cloud
optical depth matter but also the location of the cloud. One can imagine a two
layer, thin cirrus above thick cirrus, system where the optical depth of both cir-
rus clouds are essentially reduced because assigning a single optical thickness will
redistribute the radiative properties of each cloud layer to a single layer. This
could be an explanation for our LZRH\textsubscript{alt} being much lower in the atmosphere
than the LZRH\textsubscript{alt} derived in [CLP05]. An additional, and simpler explanation, is
that our results are aliased to conditions around 13:30 whereas the computations
in [CLP05] better captures the true diurnal cycle. Finally, our cloud distribution
does not include radiative cooling from the tops of deep convective clouds below
thin cirrus clouds [HHF01]. We believe this will not impact our results too much
as the cloud frequency of occurrence of deep convective cores are much smaller
than the cloud extent of the thick and thin cirrus clouds (see Figure 4.1).

The “All-Sky” curve (Figure 5.4b) is very similar to the “All-Sky” (No Radar)
curve in the TTL and more consistent with the “All-Sky” (No Lidar) curve from
the UT down into the free troposphere. From this we can easily determine that
the thin-cirrus clouds contributes 75\% (30 hPa/40 hPa) of the heating responsible
for moving LZRH\textsubscript{clear} while the thick cirrus observed by CloudSat contributes the
other 25\% (10 hPa/40 hPa), affirming that optically thin cirrus are significant
modifiers of the tropical radiative balance. This result may seem counter-intuitive
as one might expect the cooling from the cloud tops of the thick cirrus [RR89]
(also see CloudSat LW dashed blue curve in Figure 5.4b) to compensate for the
thin cirrus induced heating. This does partially occur around the base of the
TTL at $\sim$150 hPa. However, the prevalence of thin cirrus at higher altitudes
contributes strongly to the longwave heating where the CloudSat longwave cool-
ing contribution falls off.
5.4 Discussion

We have constructed an integrated atmospheric and cloud profiling dataset that spans the entire troposphere and stratosphere. From co-locating AIRS and MLS temperature and humidity (specific and relative) we showed that cloudy conditions correspond to higher RH conditions where below the TTL H$_2$O dominates RH while in the TTL temperature dominates RH. In conjunction with a realistic surface albedo model and reasonable parameterizations for cloud microphysical properties, we were able to determine the relative contributions of optically thick and thin clouds to the tropical radiative balance. In the zonal mean, the LZRH$_{\text{clear}}$ does not have much structure meridionally. However, the inclusion of clouds showed that the LZRH$_{\text{all}}$ did have a strong meridional structure where the lowest (in altitude) location of LZRH$_{\text{all}}$ corresponded to the deep tropics where thick and thin cirrus are ubiquitous.

Compositing all profiles in the tropics subtracting various cloud regimes from the “All-Sky” composites revealed that both thick and thin cirrus regimes contribute significantly to the heating in the UT and TTL. However, it was shown that the thin cirrus has a factor of 3 larger impact on the LZRH$_{\text{clear}}$ than the thick cirrus clouds resulting in a shift of LZRH$_{\text{clear}}$ from 140 hPa to LZRH$_{\text{all}} = 180$ hPa. This large shift in LZRH$_{\text{clear}}$ is a result of the prevalence of optically thin cirrus in the TTL, indicating the location of the clouds are important for determining their radiative impacts on the atmospheric column. We do note that the choice of IWC for the cirrus do matter substantially. Reducing IWC for the thin cirrus case by a factor of 5 would reduce $D_e$ to about 2.7 $\mu$m (using equation 5.1, would essentially produce a LZRH$_{\text{all}}$ similar to LZRH$_{\text{clear}}$. However, [LGY08] showed that the range of $D_e$ from in-situ measurements fall between $\sim$20-200 $\mu$m. Thus, our selections of IWC are reasonable. If anything, since our IWC yielded
particle sizes that were on the lower end of the 20-200 μm range, the results of this study may actually underestimate the impact of both cloud regimes.

Our results do have implications on interpreting the main sources of upwelling in the tropical UT. Since thin cirrus do have a substantial impact on the radiative heating in the TTL, capturing their impacts in climate models is crucial to capture the radiative impact of these clouds on stratospheric H₂O, which also impacts the TTL radiative budget [SRP10]. Finally, better characterizing the distribution of thin cirrus is also important because they can also have a more direct impact on stratospheric H₂O through dehydrating horizontally advected air in the TTL [HG01].
CHAPTER 6

Conclusion

“From small beginnings come great things.” – A Proverb

As described in the introduction, the processes that govern the interannual variations of UTLS temperature and H\textsubscript{2}O are not well understood, especially in relation to joint roles of the ENSO and QBO. The current generation of satellite sounders and active profilers on the A-Train, including AIRS, MLS, CloudSat, and CALIPSO offer unprecedented vertical information and sensitivity, which may allow us to gain new insights into these processes. The data, in concert, enables us to explore the joint distributions of temperature, H\textsubscript{2}O, cloud amount, and the radiative impacts in a self-consistent manner.

In this work we developed a integrated observational dataset of temperature, H\textsubscript{2}O, and vertically resolved clouds to characterize the impacts of the ENSO and QBO on the tropical atmosphere. While AIRS provides high quality global temperature soundings from the surface through the stratosphere, it cannot provide high fidelity H\textsubscript{2}O soundings for pressures less than \(\sim \)250 hPa due to a lack of sensitivity, in the infrared, at these altitudes. MLS provides high quality H\textsubscript{2}O down to \(\sim \)316 hPa with overlapping sensitivity with AIRS. Thus, we characterized the AIRS and MLS averaging kernels in order to develop a method to splice these to soundings into a single self-consistent integrated H\textsubscript{2}O dataset. We suc-
cessfully produced smooth profiles that did not deviate from either instruments interpretation of the atmospheric state within their uncertainties.

This enabled us to develop a ∼6 year time-record of temperature, H$_2$O, and RH. We found that the combined AIRS and MLS dataset captures well-established features of the tropics. Specifically, the RH distribution tends to follow H$_2$O until the level of neutral buoyancy, and above H$_2$O follows temperature. Furthermore, we estimated the mean H$_2$O concentrations and saturation mixing ratio and found these values to be consistent with a number of previous studies (e.g. [Des98, ZGZ01a]). This gave us confidence that A-Train measurements provided high quality measurements of temperature and H$_2$O. From these measurements, we estimated the year-to-year (de-seasonalized) variations, i.e. the interannual variability, of these quantities and were able to determine both the ENSO and QBO impacts on temperature and H$_2$O. Both of these tropical modes have distinct impacts on the tropical tropopause region temperature and H$_2$O structure with the QBO dominating in the zonal mean and ENSO contributing asymmetrically (in longitude) to the temperature and H$_2$O distribution. When the QBO and ENSO are in phase the ENSO enhances (reduces) the zonal dehydration between La Niña (El Niño) years. More specifically, the ENSO supports the QBO in enhancing temperature and H$_2$O anomalies in the TWP while reducing (or changing the sign) the anomalies in the TCP. However, the ENSO supports anomaly enhancements in the TCP, while reducing them in the TWP, when the ENSO and QBO are out of phase. The migration of convection, due to changes in the Walker Circulation, is one of the main processes that breaks the zonal symmetry of the QBO. From the time-record, we were able to find that when the ENSO and QBO are out of phase the zonal mean H$_2$O distribution is no longer just a function of processes in the TWP, but also processes that impact the TCP. It was found that during this out of phase period, the stratospheric tape-
recorder was also impacted. This suggests that the relative phase of the ENSO and QBO can impact stratospheric H$_2$O amount, which is radiatively important in this region of the atmosphere [SRP10].

Because the ENSO and QBO have such distinct signatures in the tropical tropopause layer (TTL), we hypothesized that the total tropical cloud distribution should also be impacted. From the combined CloudSat and CALIPSO observations of vertically resolved cloud layers, we found that processes linked to the ENSO and QBO do impact the tropical cloud frequency of occurrence (CFO). The ENSO impact on CFO primarily shows up as a quadrupole distribution of CFO anomalies ($\Delta CFO$) between the TWP and TCP. During El Niño (La Niña), the TCP (TWP) experiences positive (negative) $\Delta CFO$ through the much of the troposphere, and negative (positive) $\Delta CFO$ in the boundary layer, with the $\Delta CFO$ patterns mimicking H$_2$O up to $\sim$15 km and temperature above. The temperature patterns above $\sim$15 km are consistent with eastward propagating Kelvin waves induced by persistent tropical convective heating. Furthermore, the strength of each ENSO cycle is discernible by the cloud observations.

The zonally symmetric distribution of cloud amount showed coherent patterns that qualitatively followed both the ONI and QBOI. Further analyses of $\Delta CFO$ showed that the only statistically significant correlations of $\Delta CFO$ were with the QBOI. Although the correlations are significant, they are not conclusive given the brevity of the cloud record. From the $\Delta CFO$ distribution we identified the largest variations in cloud amount resulted from changes in the thin cirrus, i.e. clouds only observed by CALIPSO. Because these clouds also contribute the most to the mean tropical CFO – about a factor of 3 greater than all other tropical high clouds – their interannual variations may also impact the tropical radiative structure, specifically the level of zero radiative heating (LZRH). To
test which high clouds are radiatively most important, we computed heating rate profiles globally for a large fraction of pixels with thick and thin cirrus. We found that the cloud distribution were consistent with the observed temperature and humidity structure. We also found that while both thick and thin cirrus clouds heat the atmospheric column, changing the clear-sky LZRH\textsubscript{clear}, the thin cirrus have the greatest impacts on the LZRH. We estimated that the thin-cirrus shifted the LZRH by $\sim$1.5 km due to its amount and location. This suggests that ENSO and QBO induced changes in the tropical thin cirrus budget can impact the tropical radiative balance which is crucial for understanding the transport pathways of trace gases into the stratosphere. A future analysis will involve computing the interannual changes of radiative heating and characterizing how both the ENSO and QBO impact the tropical radiative balance.

From our analysis we have determined that the ENSO and QBO does impact the TTL temperature, H\textsubscript{2}O, and cloud structure. With this new integrated dataset we can now not only quantify the interannual impacts on the tropical atmosphere, but also the mid-latitudes. From a process standpoint, we can begin to explore the connection between the TTL cloud and H\textsubscript{2}O structure. The insights we gain will not only help us understand why the stratosphere is so dry, but also may help us understand possible cloud feedback mechanisms. We do note that the satellite measurements, especially for temperature and H\textsubscript{2}O, are limited to vertical resolutions of $\sim$2-3 km. It is unclear how this may or may not impact our understanding of the processes that govern the TTL distribution of H\textsubscript{2}O. Future work will involve analyzing high resolution in-situ measurements of H\textsubscript{2}O and clouds from aircrafts including the up-coming Airborne Tropical Tropopause Experiment (ATTREX) mission designed specifically to measure temperature, H\textsubscript{2}O, cloud ice microphysical properties, and other chemical tracers to better characterize the physical processes responsible for the observed TTL structure.
This, in conjunction with the globally sampled A-Train measurements, will provide insight into the fidelity of satellite observations of the relationship between clouds and H$_2$O.
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