The Observed Sensitivity of High Clouds to Mean Surface Temperature Anomalies in the Tropics

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Abstract. Cloud feedback represents the source of largest uncertainty in projections of future warming, and observational constraints on both the sign and magnitude of the feedback remain lacking. In this study we use observations from a suite of satellite instruments to assess the sensitivity of tropical high clouds to tropical mean surface temperature anomalies associated with interannual variability. We relate cloud changes to a physical governing mechanism that is sensitive to the vertical structure of warming, removing the ambiguity associated with simple regressions of clouds on surface temperature alone. Specifically, we demonstrate that the mean and interannual variability in both the peak level and fractional coverage of tropical high clouds as measured by CloudSat, MODIS, AIRS, and ISCCP are well-diagnosed by upper tropospheric convergence computed from the mass and energy budget of the clear-sky atmosphere. Similar to clouds in global warming simulations, observed high clouds rise and exhibit a reduction in coverage when the Tropics warm. In observations, tropical warming is accompanied by a reduction in cloud fraction at lower levels that exceeds the increase in cloud fraction at higher levels, causing absorbed solar radiation to increase more than does outgoing longwave radiation, implying a positive net cloud feedback in response to ENSO. The results suggest that the convergence metric based on simple mass and energy budget constraints may be a powerful evaluation tool for understanding observed cloud changes and for assessing the realism of modeled cloud changes in response to a variety of forcings.
1. Introduction

The role of cloud-induced changes in top of atmosphere radiative fluxes as a feedback on a warming climate is a subject of great debate and uncertainty (e.g., Bony et al. [2006]). The magnitude of cloud feedback is generally positive in global climate models (GCMs), but exhibits considerable inter-model spread that arises primarily from the spread in short-wave (SW) cloud feedbacks that can be attributed to the wide range of modeled responses of subtropical marine boundary layer clouds (e.g., Bony and Dufresne [2005]). Although most of the spread in estimates of climate sensitivity from GCMs can be attributed to the inter-model variance in SW cloud feedback, Zelinka and Hartmann [2010], hereafter ZH10, showed that the longwave (LW) cloud feedback is robustly positive in twelve GCMs in the in the the World Climate Research Programme’s (WCRP’s) Coupled Model Intercomparison Project phase 3 (CMIP3) multi-model dataset. They estimated that the tendency for tropical high clouds to rise as the climate warms contributes 0.5 W m\(^{-2}\) K\(^{-1}\) to the global mean LW cloud feedback, making it robustly positive. Furthermore, they demonstrated that the radiatively-driven clear-sky diabatic convergence, whose peak corresponds closely with the level of peak convective detrainment and abundant high cloudiness, provides a useful tool for accurately diagnosing the upward shift in cloud fraction in all of the CMIP3 GCMs analyzed. The robust nature of the positive LW cloud feedback, therefore, arises simply as a fundamental result of the approximate radiative-convective equilibrium that any model must maintain in the Tropics regardless of the details of its convection scheme. However, unlike the isothermal upward shift of high clouds expected from the fixed anvil temperature (FAT) hypothesis of Hartmann and Larson [2002], ZH10 found that high
clouds warmed slightly in the SRES A2 simulations, although the relationship between clear-sky radiatively-driven convergence and cloud fraction was followed fairly closely, a feature they referred to as the proportionately higher anvil temperature (PHAT).

Considering that climate models include a convective parameterization that adjusts toward radiative-convective equilibrium, it is perhaps not surprising that the tropical mass and energy balance is effective at diagnosing the altitude of peak modeled high cloud coverage. Kubar et al. [2007] demonstrated the close relationship between clear sky convergence and high cloud fraction measured by MODIS in three regions of the Pacific ITCZ, however, indicating that the real atmosphere is also reasonably explained by an assumption of radiative-convective balance. Still, it is unclear on what spatial and temporal scales the constraints imposed by this balance are most applicable. Convection responds quickly to variations in temperature and humidity in its near vicinity, and by moistening the near environment convection can improve the conditions for its own existence, aggregate, and achieve higher altitudes. Furthermore, organized large-scale motion can cool the air adiabatically and provide for deeper convection. For convection to continue, however, radiation must destabilize the vertical profile of temperature diabatically, and this becomes an inefficient process at low temperatures in the upper tropical troposphere, where the saturation vapor pressure is very low and water vapor becomes a less effective emitter (Hartmann et al. [2001a]).

In contrast to the lack of sensitivity of tropical high cloud top temperatures to surface temperature ($T_{sfc}$) changes expected from the FAT hypothesis, Chae and Sherwood [2010] showed that cloud top temperatures observed by the Multiangle Imaging Spectroradiometer (MISR) exhibit appreciable seasonal fluctuations (∼5K) that are associated with lapse
rate changes in the upper troposphere. This study looked at a limited domain rather than at the cloud properties of the entire Tropics; thus it remains unclear whether tropical cloud fields exhibit compensatory changes in structure that result in minimal changes when integrated over the entire Tropics. Indeed, Xu et al. [2005], Xu et al. [2007] and Eitzen et al. [2009] demonstrated that the distribution of tropical high cloud top temperatures remains qualitatively unchanged across significantly different SST distributions. These studies did not attempt to show consistency between high clouds and the radiatively-driven clear-sky convergence, however, which would have put the cloud response on more solid theoretical footing. Thus, the small body of literature that exists on the distribution of cloud top temperatures is equivocal with regard to FAT, and has not yet been assessed in light of PHAT, which accounts for changes in static stability that affect cloud top temperature.

The change in the altitude of peak cloudiness with underlying temperature is not the only aspect of high cloud changes with relevance for cloud feedback that has been investigated observationally. Lindzen et al. [2001] presented results that implied a decrease in cirrus detrainment from deep convective cores as SSTs increase (the adaptive iris hypothesis), which the authors hypothesized was due to increasing precipitation efficiency in convection over warmer waters. The vigorous debate in the literature that continues to the present casts doubt on the robustness of the results (e.g., Harrison [2002], Hartmann and Michelsen [2002a], Hartmann and Michelsen [2002b], Del Genio and Kovari [2002], Lin et al. [2002], Chambers et al. [2002], Rapp et al. [2005], Lin et al. [2006], Su et al. [2008]). However, neither the iris paper nor the responses it spawned have utilized the clear-sky diagnostics that operate on a gross Tropics-wide scale to explain changes in high clouds in observations. Rather, all have attempted to link cloud properties to the
underlying SSTs, which we argue offers a weak constraint on high cloud properties and
thus makes it difficult to draw conclusions relevant to a warming climate.

In this study we assess the degree to which the distribution of tropical cloud tops as
measured by a suite of satellite instruments changes in a manner consistent with that
predicted by the clear-sky energy budget as the Tropics warm and cool. We focus pri-
marily on the period September 2002 through July 2010, for which a wealth of satellite
information is available from A-Train instruments, but also make use of the longer cloud
record from ISCCP that extends back to July 1983. Additionally, information about the
height and optical depth of clouds from MODIS will be used in conjunction with a ra-
diative transfer model to estimate of the impact of interannual cloud fluctuations on the
TOA energy budget.

We wish to stress that our analysis is not predicated on the assumption that cloud
fluctuations associated with El Niño - Southern Oscillation (ENSO) are surrogates for
those accompanying global warming forced by increased greenhouse gas concentrations.
Rather, we demonstrate that a metric based on fundamental principles of saturation
vapor pressure, radiative transfer, and mass and energy balance accurately diagnoses the
vertical structure of tropical high clouds and its fluctuations observed in nature, just as
it does in global warming simulations of GCMs (ZH10) and in cloud resolving model
experiments of Kuang and Hartmann [2007]. The observational results presented here
reinforce the value of this diagnostic tool for understanding the varied response of high
clouds to different forcings operating across timescales and for evaluating modeled tropical
high cloud changes and their implied feedbacks.
2. Data

We make use of data from several instruments onboard satellites flying as part of the A-Train constellation. Because the satellites ascend (descend) across the equator at approximately 1:30 PM (AM) local time, the instruments are observing essentially co-located scenes. We also make use of a long record of cloud top properties from the International Satellite Cloud Climatology Project (ISCCP), TOA fluxes from CERES on Terra, and global $T_{sfc}$ from the Hadley Center (HadCRUT3v).

2.1. Atmospheric Infrared Sounder (AIRS)

AIRS is actually a suite of instruments – a hyperspectral infrared instrument (i.e., AIRS), the Advanced Microwave Sounding Unit (AMSU)-A, and a visible/near-IR sensor (Aumann et al. [2003]). The AIRS retrieval algorithm makes use of a novel cloud clearing technique that exploits the relative insensitivity of microwave temperature measurements to the presence of clouds, allowing retrievals to be made in the presence of up to 70% cloud cover (Aumann et al. [2003], Susskind et al. [2003]).

Cloud fraction reported by AIRS is actually the product of geometric cloud fractional coverage and its emissivity at 11 $\mu$m. The effective cloud fraction and cloud top pressure are derived by comparison of the observed AIRS radiance with a computed cloud radiance calculated using the surface skin temperature and atmospheric temperature-moisture-ozone profile retrieved from the clear column radiances, along with an assumed spectrally-constant cloud emissivity of 1 (Susskind et al. [2003]). The AIRS instrument saturates at an IR optical depth greater than about 5 (Huang et al. [2004]). Kahn et al. [2007] have shown that – with the exception of very thin tropopause cirrus to which AIRS is
insensitive – AIRS and Microwave Limb Sounder retrievals of cloud top pressures agree very well.

We use retrievals of cloud fraction, water vapor mixing ratio, temperature, geopotential height, and total- and clear-sky OLR from the AIRS version 5, level 3 daily gridded product (AIRX3STD) between September 2002 and July 2010. Temperature and humidity profiles are used as input to the Fu-Liou radiative transfer code (Fu and Liou [1992]) to calculate the radiative cooling rates that are used in determining the clear-sky convergence profile as explained below. Geopotential heights are used to convert co-located CloudSat retrievals to a common pressure grid.

2.2. Microwave Limb Sounder (MLS)

We make use of temperature and water vapor mixing ratio measurements from MLS onboard the Aura satellite. The instrument and its measurement technique are described in detail in Waters et al. [2006]. MLS scans downward through the atmospheric limb to retrieve profiles by observing millimeter and submillimeter wavelength thermal emission in the instrument’s field of view. Measurements are made simultaneously and continuously during both night and day, and are relatively insensitive to aerosol or thin high clouds.

Atmospheric temperature and pressure are retrieved based on emission from the spectral lines of molecular oxygen at 118 and 239 GHz. The vertical resolution is \( \sim 13 \) km at 0.001 hPa, increasing to 6 km at 316 hPa, and to 3 km at 31.6 hPa (Schwartz [2008]). Temperature precision is \( \sim 3 \) K at 0.001 hPa, increasing to 1 K or better from 3.16 hPa to 316 hPa (Schwartz [2008]); thus, only temperatures between 0.001 and 316 hPa are recommended for scientific use.
The water vapor product is taken from the 190 GHz retrieval and has vertical resolution of 3.5 km between 4.5 hPa and 147 hPa increasing to 1.5 km at 316 hPa. Between 316 and 147 hPa, MLS v2.2 has an accuracy better than 25% for water vapor mixing ratios less than 500 ppmv ([2007]). The precision increases from 25% at 147 hPa to 65% at 316 hPa ([2007]).

We use MLS version 2.2 (v2.2) Level 2 temperature and humidity data to supplement the AIRS profiles in the upper troposphere / lower stratosphere for the period August 2004 through July 2010. Both the temperature and water vapor data from MLS were screened for all flags described in the data quality and description document.

2.3. Moderate Resolution Imaging Spectroradiometer (MODIS)

MODIS is a whiskbroom-scanning radiometer with 36 channels between 0.415 and 14.235 µm. The cloud detection algorithm uses up to 20 channels to create a 1 km resolution cloud mask, which is essentially a measure of the confidence that the field of view is clear ([2003]). Assuming at least 4 of the 25 pixels are flagged as either probably cloudy or cloudy, cloud top pressure (CTP) and the effective cloud amount on a 5x5 pixel (5 km at nadir) scene are inferred using CO₂ slicing within the 15-µm absorption band. The 0.65-, 0.86-, and 1.2-µm bands (along with inferences about cloud phase) are used to retrieve optical thickness (τ), but these data are restricted to daytime observations.

We make use of cloud fraction, CTP and τ from the 5 km level 2 Joint product over the period September 2002 to July 2010. After removing retrievals in which the cloud mask is undetermined or affected by sun glint, we calculate CTP-τ joint histograms of cloud
fraction at 1° horizontal resolution, with 50 hPa-wide CTP bins between 50 and 1000 hPa and the same \( \tau \) bins of Kubar et al. [2007].

2.4. CloudSat

The primary instrument on CloudSat is the 94 GHz nadir-pointing Cloud Profiling Radar (CPR) that measures the power backscattered from cloud particles as a function of distance from the radar (Stephens et al. [2002]). It provides an instantaneous horizontal footprint of approximately 1.3 km across-track width by 1.7 km along-track length every 1.1 km along its track. Each profile contains 125 bins, each approximately 240 m thick.

We make use of the 2B-GEOPROF (Cloud Geometrical Profile) Release 4 Version 011 product between June 2006 and July 2010 and process it onto a 1° horizontal and 250 m vertical grid. The GEOPROF algorithm estimates the radar reflectivity factor for those vertical levels in which the CPR receives a significant echo (Stephens et al. [2002]). This product is referred to as the cloud mask because it indicates the presence of a cloud inside a CPR bin. Increasing cloud mask values between 20-40 represent clouds with lower chance of a being a false detection; the threshold of 20 carries a 5% false detection rate (Marchand et al. [2008]).

We keep only the profiles for which the quality flags indicate good data and create a binary cloud mask containing ones where the GEOPROF cloud mask value is greater than or equal to 20. From this binary cloud mask, we compute a binary profile of cloud tops by simply locating every vertical bin with a value of one directly below one with a value of zero. Finally, to facilitate comparison with other datasets, we interpolate the CloudSat data from its native geometric height grid to a pressure grid using co-located
AIRS retrievals of geopotential height and to a temperature grid using the combined
AIRS-MLS temperatures described below.

2.5. International Satellite Cloud Climatology Project (ISCCP)

ISCCP provides global cloud data derived from radiance-calibrated infrared and visible
radiances obtained from polar orbiting and geostationary satellites. We make use of
the ISCCP-D1 cloud product, which is a 3-hourly global dataset on an equal-area grid,
providing cloud fractions as joint functions of seven cloud top pressure bins and six optical
depth bins. We use data from ISCCP’s entire current period of record (July 1983 - June
2008). For detailed description of the D1 dataset, refer to Rossow and Schiffer [1999].

Briefly, scenes are classified as cloudy if the IR or VIS radian ce in the 4-7 km field of view
differs from the clear-sky value by more than the detection threshold. Cloud fraction is
reported for the larger 280 km grid as the fraction of all 4-7 km pixels containing clouds.

In the D-series datasets, biases in detectable cloud amounts are about 0.05, except in the
summertime polar regions where the bias may be about 0.10.

Cloud top temperature is computed for each cloudy scene by comparing the IR radian ce
observed with that computed from a radiative transfer model. Then, cloud top pressure
is determined using a temperature profile from the TIROS Operational Vertical Sounder.

Biases in cloud-top temperatures are less than $\sim$2 K for lower-level clouds and less than $\sim$4
K for optically thin, upper-level clouds, except when they occur over lower-level clouds
(Rossow and Schiffer [1999]). Optical thickness is computed for each cloudy scene by
comparing the VIS radian ce observed with that computed from a radiative transfer model.

Differences in actual and modeled cloud microphysical properties lead to biases in the
ISCCP cloud optical depths of -4% over ocean and +2% over land (Han et al. [1994]).
2.6. Clouds and the Earth’s Radiant Energy System (CERES)

We make use of TOA clear- and total-sky LW and SW fluxes from several CERES products (Wielicki et al. [1996]) from sensors onboard both Aqua and Terra spacecrafts. These fluxes are used to compute monthly mean values of SW and LW cloud forcing.

The CERES Edition 2.5 “lite” dataset uses Edition 3 calibration and Edition 2 processing to remove all known CERES instrument artifacts. We use clear- and total-sky TOA fluxes from the two 1° gridded monthly mean datasets, the SSF-lite Edition 2.5-Beta and the SYN-lite Edition 2.5-Beta, which cover the period March 2000 through February 2010. The SYN product provides diurnally averaged fluxes by making use of 3-hourly geostationary data to estimate the flux in between CERES measurements during which time meteorology may change (Young et al. [1998]). In contrast, SSF uses ERBE temporal interpolation which assumes constant meteorology between CERES measurements and does not account for regional diurnal changes in flux and cloud properties. Both “lite” products are derived from measurements by CERES on Terra.

The ERBE-like Monthly Regional Averages (ES-9) product contains daily averages of instantaneous SW and LW TOA fluxes computed using the available hourly data, scene identification data, and diurnal models in an algorithm similar to that used for the Earth Radiation Budget Experiment (ERBE). Fluxes are given for both clear-sky and total-sky scenes, though over large regions of the Tropics the clear-sky data is missing, even on monthly timescales if the region is particularly cloudy. We use the ES-9 flight module (FM) 3 Edition 2 dataset from CERES on Aqua from the period July 2002 through Feb 2010.
The Energy Balanced and Filled (EBAF) product is recommended for estimating the Earth’s global mean energy budget, as net TOA fluxes are constrained by the ocean heat storage term. It is derived from the SYN product which uses the diurnally complete temporal averaging and the CERES angular directional models, and the algorithm spatially interpolates clear-sky fluxes to fill the gaps in non-observed regions. We use the entire current EBAF Edition 1a dataset from CERES on Terra, (March 2000 - Oct 2005).

The SRBAVG product contains gridded monthly mean clear- and total-sky TOA LW and SW GEO-interpolated fluxes at 1° resolution. We use the entire current SRBAVG1 FM3 Edition 2a dataset from CERES on Aqua (July 2002 - Oct 2005).

2.7. HadCRUT3v

We make use of the globally gridded HadCRUT3v $T_{sfc}$ dataset provided by the Hadley Center (Brohan et al. [2006]). 4349 land stations, along with marine data from in situ ship and buoy observations are used to construct the dataset. Sea surface temperatures are assumed to be a good surrogate for surface air temperatures over the ocean. HadCRUT3v is the variance-adjusted version of the HadCRUT3 dataset, meaning that each grid box’s anomalies are adjusted to account for a changing number of observing sites over the period of record. The data cover the period from January 1850 to the present, but we only utilize the portion that overlaps with the satellite records used here.

3. Methodology

3.1. Combining AIRS and MLS temperature and humidity profiles

As shown in Kubar et al. [2007], the profile of $conv$ and its fluctuations are sensitive to the structure of upper tropospheric - lower stratospheric (UTLS) temperature
and humidity profiles. Unfortunately, the UTLS region is a particularly difficult area of the atmosphere to measure these quantities accurately (Kley et al. [2000], Soden et al. [2004a]). In Figure 1 we show the area-weighted all-sky tropical mean temperature and humidity profiles in the UTLS region measured by AIRS and MLS, along with GPS occultation measurements of temperature from Constellation Observing System for Meteorology, Ionosphere, and Climate (COSMIC) (Anthes [2008]). (GPS mixing ratios are primarily model-generated in the UTLS, so they are not shown.) In general, the temperature profiles are in agreement at all levels, though AIRS places the cold-point tropopause somewhat lower in the atmosphere than do the other datasets. At pressures greater than about 150 hPa, AIRS and MLS mixing ratios are in good agreement as was shown in Read [2007], but AIRS is significantly drier than MLS above this level.

We have chosen to combine the AIRS and MLS temperature and humidity profiles in such a manner that each dataset is used where it is most reliable, with a transition pressure of 200 hPa. The transition from AIRS to MLS data is done by giving increasing weight to the MLS data relative to the AIRS data as the 200 hPa level is approached from below.

3.2. Computation of clear-sky radiative cooling, diabatic subsidence, and diabatic convergence

We follow the procedure described in Section 3 of ZH10 to compute profiles of clear-sky radiative cooling \( Q_R \), diabatic subsidence \( \omega \), and clear-sky diabatic convergence \( \text{conv} \). Briefly, we assume that clear-sky radiative cooling (calculated using the Fu-Liou radiation code, with zonal- and monthly-mean combined AIRS-MLS profiles of temperature and humidity as input) is exactly balanced by warming due to diabatic subsidence:

\[
\omega = \frac{Q_R}{\sigma}.
\]
σ is the static stability, which can be written

\[ \sigma = \frac{\kappa T}{p} - \frac{\partial T}{\partial p}, \] (2)

where \( \kappa = R_d/c_p \), \( R_d \) is the gas constant for dry air, and \( c_p \) is the specific heat of air at constant pressure. Assuming mass continuity, the profile of \( \text{conv} \) in the clear-sky region is calculated by

\[ \text{conv} \equiv -\nabla_H \cdot U = \frac{\partial \omega}{\partial p}. \] (3)

Assuming a closed mass budget between convective and nonconvective regions, the rate of convergence into the clear-sky region is equivalent to the divergence out of the convective region. The peak in this radiatively-driven mass divergence is a marker for the top of the rapidly mixed troposphere and is expected to be co-located with convectively-detrained anvil clouds.

### 3.3. Regressions on Tropical Mean Surface Temperature Anomalies

For each variable we first compute area-weighted tropical mean monthly means over their period of record. We then compute anomalies of each monthly mean datapoint from this average annual cycle of monthly data. Sensitivities to tropical mean surface temperature (\( \bar{T}_{\text{sfc}} \)) are calculated as regression coefficients between each variable and \( \bar{T}_{\text{sfc}} \) anomalies. Estimates of the uncertainty in the derived regression slopes are computed using a bootstrapping method in which the predictand is re-sampled 10000 times to compute a distribution of possible regression coefficients. We take the 2σ spread in regression slopes as representing the 95% confidence interval surrounding each regression slope. Slopes for which the 2σ range includes zero are considered statistically insignificant at the 95% confidence level.
4. Results

4.1. Consistency Between High Cloud Fraction and Diabatic Convergence

In Figure 2 we show the tropical mean combined AIRS-MLS mixing ratio and temperature profiles, the radiative cooling calculated with the Fu-Liou radiative transfer code, and the static stability, diabatic subsidence, and diabatic convergence calculated by Equations 1 - 3. Tropical temperatures approximately follow the moist adiabat (Xu and Emanuel [1989]) at pressures greater than about 250 hPa, above which level the temperature profile is more stable than the moist adiabat (c.f., Figure 1 of Mapes [2001]).

Mixing ratios decrease exponentially with decreasing pressure due to the exponential dependence of saturation vapor pressure on temperature and the decrease in temperature with decreasing pressure. The $Q_R$ profile exhibits a cooling of about 1.5 K dy$^{-1}$ that is nearly constant with pressure up to about 250 hPa, above which the profile decreases dramatically with decreasing pressure to a level of zero radiative heating at around 100 hPa. This decrease in $Q_R$ is related to the extremely small concentrations of water vapor present in the upper troposphere. Static stability, given by Equation 2, is small and roughly constant with pressure up to about 200 hPa, above which point the combination of a decreasing lapse rate and the inverse-pressure dependence cause a large increase with decreasing pressure. The implied diabatic subsidence that is necessary to balance $Q_R$, given by Equation 1, is relatively constant at 30 hPa dy$^{-1}$ at pressures greater than 250 hPa, then decreases rapidly with decreasing pressure, reaching a value of zero at about 100 hPa (where $Q_R$ is also zero). The rapid decrease of diabatic subsidence is related to both the rapid decrease of $Q_R$ and the rapid increase of $\sigma$ in the upper troposphere. The
implied upper tropospheric $\text{conv}$, given by Equation 3, exhibits a large peak at 200 hPa where the decrease of diabatic subsidence with decreasing pressure is most dramatic.

Profiles of tropical mean cloud top frequency of occurrence from CloudSat as well as cloud top fraction from AIRS, MODIS, and ISCCP are shown in Figure 3. MODIS cloud fractions are plotted at the geometric mean pressure of the cloud top pressure bins. The MODIS cloud fraction profile includes all clouds, regardless of their optical depths, and includes retrievals for which cloud top pressure but not cloud optical depth is retrieved. In the case of ISCCP, only clouds with optical depths exceeding 1.3 are included because the ISCCP retrieval algorithm places a questionably large fraction of clouds into the highest, thinnest bin of the histogram. Marchand et al. [2010] explain that the ISCCP algorithm is generally unable to accurately determine the optical depth if it detects a cloud based on the IR threshold but the visible reflectance is very close to the expected clear-sky value. In this situation, the ISCCP algorithm assigns the cloud top temperature to the expected tropopause temperature minus 5 K, with a resulting cloud top pressure near that of the tropopause. Removing this cloud type results in a tropical mean ISCCP cloud top profile that is similar to that derived by the other instruments, with a peak in the 180-310 hPa bin. Overlaid as dashed lines in each figure for comparison is the $\text{conv}$ profile shown in Figure 2.

It is important to bear in mind that $\text{conv}$ is a measure of the net convergence into the clear-sky regions that is required by the net diabatic tropical overturning. Thus one should not interpret $\text{conv}$ as a quantity to which cloud fraction should be proportional at every height. (If this were the case, cloud fraction would be zero or even negative throughout most of the lower and middle troposphere.) Rather, the integrated $\text{conv}$ is a measure
of the net mass flux in the divergent circulation of the Tropics, the upper tropospheric branch of which is associated with detrainment from deep convection and its attendant anvil cloud coverage.

As was the case in the observational study of Kubar et al. [2007], the cloud resolving model study of Kuang and Hartmann [2007], and in the GCM study of ZH10, the peak in the profile of conv is remarkably well-correlated with the peak in the cloud profiles. The profile of CloudSat cloud tops exhibits a particularly striking similarity to the conv profile, which makes sense considering CloudSat’s sensitivity to larger ice particles, removing any strong influence from very thin cirrus that would potentially peak at a higher altitude. It also suggests that the peak in the profile of conv serves as a convenient marker for the emission level of the bulk of tropical high clouds. In general, the peaks in AIRS and MODIS cloud top fraction and CloudSat cloud top frequency of occurrence tend to lie slightly above the peak in conv.

We have separated the MODIS cloud fractions into three optical depth ranges corresponding to thin \((0 \leq \tau < 4)\), anvil \((4 \leq \tau < 32)\), and thick \((\tau \geq 32)\) clouds as in Kubar et al. [2007] (not shown). Zelinka and Hartmann [2009] verified that such separation by optical depth reasonably distinguishes between ubiquitous thin cirrus that may not be connected to deep convection, thicker anvil clouds that detrain from deep convection and spread outward from the convective core, and the thick clouds in the cores of cumulonimbus updrafts. We find that thin and anvil clouds correspond most closely to the convergence profile, whereas thick cloud fraction tends to peak at a higher level than the convergence peak. This is consistent with the results of Kubar et al. [2007], who point out that the latter cloud type is more closely associated with convective updrafts.
whereas anvils detrain from the updraft at a level consistent with peak clear sky upper
tropospheric convergence.

4.2. Tropical Mean $T_{sfc}$ Fluctuations and their Associated Cloud Anomalies

We are interested in the sensitivity of cloud fields to $T_{sfc}$ and how well this sensitivity
is diagnosed by the anomalies in $\text{conv}$. Thus we make use of the HadCRUT3v dataset of
$T_{sfc}$ over the period encompassing the satellite observations of clouds. Since we will focus
primarily on the data-rich period of the A-Train, Figure 4 shows the timeseries of $T_{sfc}$
anomalies over the period September 2002 - July 2010.

The record includes considerable month to month variability, but the dominant feature
is a notable transition from the fairly neutral conditions that prevailed until early 2007 to
a strong La Niña by the end of 2008, followed by a steady warming to a strong El Niño
by the beginning of 2010. Surface temperature anomalies associated with a tropical-mean
warming (not shown) exhibit a canonical El Niño pattern, with massive warm anomalies
in the eastern tropical Pacific straddled by cold anomalies to the north, south, and west,
and large cold anomalies in southeastern North America and midlatitude Eurasia.

Before proceeding, we assess the robustness of the temperature and moisture fluctu-
ations in the upper troposphere by comparing the COSMIC, MLS, and AIRS datasets
(Figure 5). All three datasets exhibit a pronounced warming that extends up to
about 200 hPa of between 2-2.5 K per degree of tropical mean surface warming. GPS
occultation measurements exhibit the smallest upper tropospheric amplification of the
warming signal and AIRS exhibits the largest. All three datasets also exhibit large nega-
tive temperature anomalies in the lower stratosphere that peak around $-2 \text{ K K}^{-1}$ between
50 and 65 hPa. AIRS exhibits a sharp linear decrease of temperature anomaly above 200
hPa, whereas MLS and GPS curve more gently at first and then reach a more negative value at about 65 hPa. Since MLS and GPS have better vertical resolution in the upper troposphere and lower stratosphere, and agree well with each other, they may be more correct. The exact slope of the fall-off of temperature anomalies with decreasing pressure affects the $\sigma$ anomalies in the UTLS region, which impacts the implied diabatic subsidence and convergence anomalies. Thus our assessment of convergence changes as well as our determination of cloud top temperature changes are quite sensitive to the dataset chosen.

AIRS- and MLS- observed water vapor mixing ratio sensitivities are not dramatically different from each other (note that the regression slopes from each dataset fall within the $2\sigma$ range of uncertainty of each other), which is reassuring considering that humidity fluctuations affect $Q_R$ anomalies and therefore the implied subsidence and convergence.

Near 200 hPa, where our combined product is weighted equally by both products, MLS-measured mixing ratios exhibit greater sensitivity to tropical mean temperature fluctuations than do those measured by AIRS. Furthermore, AIRS mixing ratio anomalies exhibit a less-rapid fall-off with decreasing pressure above 175 hPa compared with MLS measurements. Still, considering the myriad difficulties in measuring water vapor in the UTLS, the level of agreement in the anomalies is noteworthy.

The vertical structure of temperature and humidity fluctuations has implications for the profile of $\text{conv}$. In Figure 6 we show the sensitivity of tropical mean temperature, water vapor mixing ratio, radiative cooling, static stability, diabatic subsidence, and diabatic convergence to $T_{sfc}$. The entire troposphere up to just above 100 hPa warms in association with tropical mean warming, with a peak warming occurring at about 200 hPa. Mixing ratios increase at all pressure levels, but most dramatically between 100 and
300 hPa. This is due to the combination of the warming peak at 200 hPa and simply because even tiny absolute perturbations to the humidity profile in the upper troposphere result in large fractional increases in mixing ratio because the mean concentrations are so low. Radiative cooling anomalies mimic the humidity anomalies such that where water vapor concentrations increase, $Q_R$ also increases, as expected from the FAT hypothesis. Static stability increases slightly up to about 240 hPa, then decreases substantially at pressures below 240 hPa. This structure is primarily governed by the vertical structure of warming, which peaks at 200 hPa and fairly rapidly transitions to cooling above 100 hPa.

At pressures less than 240 hPa, the combination of enhanced $Q_R$ and reduced $\sigma$ results in an increase in diabatic subsidence. Conversely, at pressures greater than 240 hPa, the combination of enhanced $\sigma$ overcompensating for enhanced $Q_R$ results in a decrease in diabatic subsidence. Anomalously large subsidence above the level of peak convergence and anomalously small subsidence below the level of peak convergence represents a reduction in the vertical derivative of subsidence, which reduces the convergence peak. Peak enhancement of subsidence occurs at 200 hPa, resulting in anomalous convergence above 200 hPa.

In summary, in response to a 1 K increase in $T_{sfc}$, the $Q_R$ profile shifts upward in association with a warmer, more moist upper troposphere, the diabatic subsidence profile shifts upward and exhibits a slightly less-dramatic decrease with decreasing pressure, and the convergence profile shifts upward and exhibits a smaller peak value, much as it does in GCMs under greenhouse warming (ZH10).
Sensitivity of observed cloud profiles to a 1 K increase in $T_{sfc}$ are shown in Figure 7, along with the sensitivity of the $conv$ profile to warming repeated from Figure 6f. All datasets exhibit large reductions in cloud top fraction or frequency around 200-250 hPa (i.e., near the peaks in their respective mean profiles). Furthermore, cloud fractions from all datasets exhibit increases at pressures less than about 200 hPa (though positive MODIS cloud top fraction anomalies are not statistically significant). The structure of these cloud changes is well-diagnosed by the change in $conv$ profile.

The profile of anomalous cloud tops from CloudSat shows a remarkable similarity to the profile of anomalous upper tropospheric $conv$, with both having the same location of zero-crossing. At pressures less than (greater than) 160 hPa, both convergence and cloud top frequency increase (decrease). Negative cloud top occurrence anomalies are greatest at the pressure of peak mean cloud top occurrence. That anomalous cloud tops measured by CloudSat most closely track the anomalous convergence profile is very reassuring given CloudSat’s superior vertical resolution relative to the other sensors.

Separating the MODIS cloud fractions by optical depth, we find that anvil and thick clouds have a tendency to rise in association with tropical warming, whereas the dominant feature of the thin cloud fraction anomalies is a reduction at about 225 hPa (not shown). The cloud fraction anomalies are not statistically different from zero at most pressure levels, but all cloud types show a significant reduction in cloud fraction near 225 hPa. The profiles of anomalous cloud fraction correspond quite well with the anomalous $conv$ profiles, with the exception of thin clouds, which do not exhibit an increase near 150 hPa. That the anvil cloud fractions exhibit the largest fluctuations in association with tropical
warming is also consistent with the interpretation of \textit{conv}, which one would expect to be physically related to mass detrainment and therefore anvil coverage.

The upward shift and reduced peak in the convergence and cloud profiles observed here are similar to those that accompany a warming climate in GCMs (ZH10), but we demonstrate in Section 4.3 that the shift in cloud profile is accompanied by smaller changes in cloud top temperature (i.e., the response is more isothermal than in models).

What is perhaps most striking is that the cloud profiles from all datasets (except ISCCP) exhibit a decrease in cloud coverage at and below their peak level that exceed increases in cloud coverage aloft. This is consistent with the net decrease in clear-sky convergence (i.e., the large decrease in clear-sky convergence at pressures greater than 160 hPa exceeds the increase at pressures less than 160 hPa). This lends some support to an iris-like response in high cloud coverage that is directly related to the decrease in upper tropospheric clear-sky convergence as the Tropics warm. Note that the mechanism invoked for such cloud changes is quite different from that of \textit{Lindzen et al.} [2001] in that it has nothing to do with cloud microphysics, but relies only on mass and energy budget considerations that show apparent skill in predicting high cloud changes.

\textbf{4.3. FAT or PHAT?}

An important finding in ZH10 is that GCM cloud fraction profiles shift upwards, but less so than do the isotherms. This meant that the cloud-weighted temperatures increased very slightly rather than staying constant as expected from the FAT hypothesis. This non-isothermal shift in cloud profile was well-diagnosed by the shift in upper tropospheric convergence and was associated with changes in the profile of \(\sigma\). Because \(\sigma\) inversely depends on pressure (Equation 2), rising isotherms experience an increase in \(\sigma\), unless the
lapse rate increases to match the adiabatic lapse rate at those higher altitudes. In the greenhouse warming case simulated by GCMs, \( \sigma \) increased at all temperatures because the effect of rising isotherms was greater than the effect of increased upper tropospheric lapse rate.

In Figure 8 we plot the variables shown in Figure 2, but as functions of temperature. Water vapor concentrations are fundamentally limited by temperature via the Clausius-Clapeyron relation; thus the profile of mixing ratio remains nearly constant in temperature coordinates. At temperatures less than about 230 K, however, small but statistically significant moistening occurs. Because \( Q_R \) is primarily due to water vapor rotation lines in the upper troposphere, its profile is also largely unchanged when plotted as a function of temperature, though cooling is slightly enhanced at temperatures colder than 210 K where moisture increases and the lapse rate increases with tropical warming.

Unlike the case of GCM-simulated global warming in which \( \sigma \) increased significantly at every temperature (c.f., Figure 4c of ZH10), here \( \sigma \) increases only very slightly when plotted as a function of temperature. At most temperatures (exceptions being 204-207 K and 215-220 K), the perturbation profile is not statistically different from the mean profile. It is important to understand why this very different structure emerges in the observations. Taken alone, rising isotherms due to warming increase stability at all temperatures (not shown) because of the inverse pressure dependence in Equation 2. This effect is nearly equally opposed by the increase in lapse rate at temperatures colder than 240 K (not shown). In modeled warming due to increased \( CO_2 \), the former effect dominated over the latter effect, resulting in large \( \sigma \) increases at all temperatures. Here, we see that these effects nearly equally offset each other, causing a negligible change to the \( \sigma \) profile. The
differences between greenhouse warming and tropical warming associated with ENSO may be related to the fact that the heat source in the former is radiative, relatively spatially homogeneous, global in nature, and results in a larger tropical mean surface warming, whereas that of the latter is due to anomalous surface heat flux that is spatially confined to the Tropics and results in a smaller tropical mean surface warming (e.g., Lu et al. [2008]). Transport of heat out of the Tropics may proceed more efficiently during a warm phase El Niño than under greenhouse warming, resulting in a less-dramatic upper tropospheric warming and increase in $\sigma$.

In Figure 8, the subtle but competing effects of changes in the $\sigma$ and $Q_R$ profiles on the diabatic subsidence profile are apparent. Everywhere except at temperatures colder than 206 K, the slightly larger increases in $\sigma$ relative to $Q_R$ result in slight reductions in diabatic subsidence, though the change is generally statistically insignificant. At temperatures below 206 K, slightly enhanced $Q_R$ and slightly reduced $\sigma$ result in increased diabatic subsidence. In the end, a small reduction in diabatic subsidence at all but the coldest temperatures causes a reduction in the peak convergence and a slight shift of the profile towards warmer temperatures. Clearly, the conv profile is shifting upwards (Figure 6f) in such a way as to remain at nearly the same temperature (Figure 8f).

Because CloudSat provides the most highly-resolved cloud top information, we compare its anomalies with those of the convergence profile as functions of temperature in Figure 9. The two profiles are remarkably similar, with decreased convergence and cloud top coverage at all temperatures, but most dramatically between 200 and 220 K. A slight shift of both the peak in convergence and the peak cloud amount towards warmer temperatures is apparent, though it is not statistically significant. Thus, we cannot rule out a purely
isothermal (FAT-like) response of the cloud tops to tropical mean warming, but the results are suggestive of a proportionately higher temperature (PHAT-like) response, as seen in GCM simulations (ZH10). Differences in the vertical structure of warming between month-to-month fluctuations shown here and greenhouse warming in GCMs leads to subtle differences in cloud responses, but the basic constraint imposed by the clear-sky energy budget fairly accurately explains cloud changes in either case.

4.4. Radiative Impact of Observed Cloud Anomalies

In this section we assess the implications of the observed cloud fluctuations for TOA radiation in two ways. First, we use histograms of MODIS-derived cloud fraction as a joint function of optical depth and cloud top pressure, combined with histograms of overcast sky cloud radiative forcing generated using a radiative transfer model to calculate the impact of the observed cloud fraction changes on TOA radiative fluxes. Second, we compute the change in tropical mean cloud radiative forcing from AIRS and CERES broadband measurements of clear- and all-sky flux.

Other than the following differences, the procedure for computing overcast-sky cloud radiative forcing histograms is the same as in Hartmann et al. [2001b], Kubar et al. [2007], and Zelinka et al. (2011a, manuscript submitted to *J. Climate*), to which the reader is referred for the details of the procedure. We insert monthly mean AIRS temperature and humidity profiles into the Fu-Liou radiation code, along with synthetic profiles of liquid or ice water content that correspond to the cloud top pressure and optical depth at the midpoint of each MODIS histogram bin. TOA fluxes computed with and without synthetic clouds are differenced to compute the individual impact of each cloud type, resulting in an overcast-sky cloud radiative forcing histogram. We compute a LW and
SW histogram for each latitude equatorward of 30°, which is then multiplied by the cloud fraction anomaly histogram to compute the effect of cloud fraction anomalies on TOA fluxes.

In Figure 10 we show the anomalous cloud fraction histogram due to a 1 K tropical-mean temperature anomaly, along with the product of this histogram with LW, SW, and net overcast-sky histograms (not shown). Note that this multiplication is giving an estimate of the the radiation balance changes caused by clouds alone, with all other quantities held fixed. Thus, the sum of the histogram is a direct estimate of cloud feedback, but we emphasize that it is the cloud feedback in response to ENSO and not to CO₂-induced global warming.

Anvil (i.e., CTP < 450 hPa and 1 ≤ τ < 16) and thick (i.e., 150 ≤ CTP < 250 hPa and τ ≥ 16) cloud fractions exhibit large anomalies. Most dramatic are the reductions in cloud fraction at pressures greater than about 180 hPa and smaller increases in cloud fraction at pressures less than about 180 hPa. Additionally, low cloud fractions exhibit a broadening of their distribution in the vertical, as evidenced by decreases near their peak straddled above and below by large increases, but we will focus henceforth on high cloud changes and their radiative implications.

The large decrease in anvil and thick clouds at and below the level of their respective peaks and slightly smaller increase above the level of their respective peaks is apparent in both the anomalous LW and SW cloud feedback histograms. High cloud fractional changes are the dominant cause of changes in LW fluxes, whereas SW fluxes are sensitive to both high and low cloud changes. Clouds with CTP<450 hPa decrease by about 1% in association with a 1 K tropical T_{sfc} anomaly, implying a negative LW high cloud feedback
(regression slope of -1.0 W m\(^{-2}\) K\(^{-1}\)). Conversely, the broad reductions in high cloud amount result in large decreases in reflection and therefore an implied positive SW high cloud feedback (regression slope of 1.4 W m\(^{-2}\) K\(^{-1}\)). The impact of high cloud changes on SW fluxes is opposed by that of low cloud changes, but nevertheless, the SW cloud feedback is positive in association with tropical warming (regression slope of 0.9 W m\(^{-2}\) K\(^{-1}\)). In the net, high cloud feedback is positive, not primarily because of the enhanced greenhouse effect from rising cloud tops but rather because of the enhanced downwelling SW radiation from reduced high cloud coverage.

As an independent check of the sensitivities computed above, regressions of AIRS- and CERES-derived tropical mean LW and SW cloud radiative forcing (LWCF and SWCF, respectively) on \(T_{sfc}\) are provided in Table 1. Cloud radiative forcing is defined as the difference between clear-sky and all-sky upwelling fluxes at the TOA (Charlock and Ramanathan [1985]); thus anomalies in cloud forcing are not the same as anomalies in LW and SW fluxes computed using overcast-sky histograms because the former includes coincident changes in non-cloud-induced fluxes (c.f., Soden et al. [2004b]). Dessler [2010] has shown that between March 2000 and February 2010, the difference between global mean cloud radiative forcing anomalies (computed using all-sky fluxes from the CERES “lite” datasets and clear-sky fluxes from MERRA reanalysis data) and radiation anomalies induced solely by clouds (computed by adjusting the cloud forcing anomalies for non-cloud-induced flux anomalies using the radiative kernels of Soden et al. [2008]) is small, but A. Dessler (2011, personal communication) noted that non-cloud-induced LW flux anomalies can be appreciable (up to 0.3 W m\(^{-2}\)) in the Tropics in certain months.
We find that the sensitivity of tropical mean LW fluxes to surface temperature derived using the LW histogram is within the errorbars of the regression slope of $LWCF$ on $T_{sfc}$ computed from direct AIRS and CERES measurements (Table 1). The same is true for the $SWCF$ regressions, with the exception of the EBAF and SRBAVG datasets that cover shorter time periods ending prior to the most recent large La Niña to El Niño transition. In all but the ES9 dataset, positive $SWCF$ anomalies dominate over negative $LWCF$ anomalies, as is the case in the histogram-derived estimates of SW and LW flux anomalies due to high clouds only. These results are in agreement with Lin et al. [2002], who showed using the same model as Lindzen et al. [2001] but with CERES TOA fluxes that the SW “iris” dominates over the LW “iris”. It is important to bear in mind that MODIS cloud fraction anomalies in many pressure and optical depth bins are statistically insignificant and that the cloud types included in the MODIS histogram do not represent all cloud types present (e.g., Marchand et al. [2010] show that MODIS frequently does not retrieve optical depths for low broken clouds and optically thin ($\tau < 1$) high clouds; thus they will tend to be excluded from the MODIS histogram). Nevertheless, the level of agreement between the values derived using these independent techniques is noteworthy, with nearly all implying a positive tropical cloud feedback acting on interannual timescales.

5. Conclusions and Discussion

We have demonstrated in this study that the upper tropospheric diabatic convergence ($conv$) that results from the balance of radiative cooling and subsidence warming in the clear-sky Tropics provides a powerful tool for diagnosing both the vertical level and magnitude of peak tropical cloud coverage as measured by a suite of satellite-borne sensors. Furthermore, we have demonstrated that fluctuations in the profiles of tropical cloud
coverage in association with interannual variability of SST are well-diagnosed by these clear-sky constraints. Specifically, as the Tropics warm in association with ENSO, cloud fraction profiles exhibit an upward shift and reduction in peak coverage, a structure that is remarkably well-diagnosed by $\text{conv}$.

In qualitative agreement with the isothermal cloud response expected from the fixed anvil temperature hypothesis and seen in the modeling studies of Hartmann and Larson [2002] and Kuang and Hartmann [2007] and observational studies of Xu et al. [2005], Xu et al. [2007] and Eitzen et al. [2009], the cloud profile exhibits small but insignificant variations when plotted in temperature coordinates. Small increases in static stability at all temperatures result in a slight reduction in peak $\text{conv}$ and a statistically insignificant shift of the $\text{conv}$ peak towards warmer temperatures, a pattern that is mimicked in the cloud profiles and is suggestive of the PHAT-like response seen in CMIP3 GCMs (ZH10).

Finally, we have made use of CTP-$\tau$ joint histograms of cloud fraction and overcast-sky cloud forcing to estimate the effect of changing cloud distribution on TOA fluxes. MODIS observes a large decrease in anvil cloud coverage at and below its peak and a smaller increase above its peak in response to tropical warming. These high cloud fraction anomalies result in a net heating of the Tropics primarily because the overall reduction in coverage enhances SW absorption more than it enhances LW emission. The sensitivities of cloud-induced TOA flux anomalies to temperature computed with this technique are generally within the range of uncertainty of those computed for CERES- and AIRS-derived cloud radiative forcing anomalies, and nearly all estimates imply a positive tropical net cloud feedback operating on interannual timescales.
We wish to stress that the results of this chapter are not meant to suggest that radiation anomalies due to cloud changes associated with ENSO can be used as a surrogate for long-term cloud feedback due to CO$_2$-induced global warming, or that the long-term global mean SW cloud feedback is positive and LW cloud feedback is negative. Rather, we have shown that the clear-sky diabatic convergence is an effective metric for diagnosing the mean and change in amount, altitude, and temperature of peak high level cloudiness in nature. These results, in combination with those of ZH10, lend credence to the utility of this tool for understanding high cloud changes due to climate fluctuations across time scales forced by a variety of mechanisms and for evaluating the realism of high cloud changes and their implied feedbacks in models.

Dessler [2010] found no correlation between cloud feedbacks derived on short time scales and those derived on long time scales. Our work offers a possible mechanism for explaining why cloud feedbacks (at least the component arising from tropical cloud changes) operating on different timescales are uncorrelated. Upper tropospheric amplification of warming is less vertically and horizontally extensive during El Niño than in global warming simulations (c.f., Figure 2 of Lu et al. [2008]). This results in a smaller upward shift and a larger decrease in implied clear-sky upper tropospheric convergence accompanying El Niño than that accompanying global warming (not shown), and these features are mimicked in the cloud fields. Thus, the vertical structure of warming – through its impact on the clear-sky convergence profile – may determine the anomalous cloud structure that arises in response to a climate perturbation. Given that the vertical structure of warming differs considerably depending on whether it is forced radiatively (e.g., by increasing CO$_2$) or by anomalous tropical air-sea heat fluxes (e.g., ENSO), it is inevitable that tropical clouds...
will exhibit a variety of responses to a given $T_{sfc}$ anomaly. Thus, cloud feedbacks inferred from regressions of cloud properties on tropical mean surface temperature fluctuations driven by short-term variability may have little relation to the long-term cloud feedback in response to increasing greenhouse gas concentrations.

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Figure 1. Tropical mean (a) temperature from (black) COSMIC, (blue) MLS, and (red) AIRS and (b) water vapor mixing ratio from (blue) MLS and (red) AIRS. The dashed lines represent the $2\sigma$ range of monthly tropical average quantities. Note that mixing ratios are plotted on a log scale.
Figure 2. Tropical mean (a) temperature, (b) water vapor mixing ratio, (c) radiative cooling, (d) static stability, (e) diabatic subsidence, and (f) diabatic convergence. Temperature and mixing ratio retrievals are from the combination of AIRS and MLS, radiative cooling is calculated with the Fu-Liou radiative transfer code, and the other terms are calculated according to Equations 1 - 3. The dashed lines represent the $2\sigma$ range of monthly tropical average quantities. Note that mixing ratios are plotted on a log scale.
Figure 3. (blue) Tropical mean (a) cloud top frequency of occurrence from CloudSat, and cloud fraction from (b) MODIS, (c) AIRS, and (d) ISCCP. Only clouds with $\tau \geq 1.3$ are included in the ISCCP cloud fraction plot. Red lines overlain in each panel show the diabatic convergence repeated from Figure 2f. The dashed lines represent the $2\sigma$ range of monthly tropical average quantities. Note that the range of values on the upper x-axis varies from panel to panel.
Figure 4. HadCRUT3v tropical mean surface air temperature anomalies relative to the 1961-1990 mean. Sea surface temperatures are used in place of surface air temperatures over the ocean.
Figure 5. Sensitivity of tropical mean (a) temperature from (black) COSMIC, (blue) MLS, and (red) AIRS and (b) water vapor mixing ratio from (blue) MLS and (red) AIRS. Sensitivity profiles are computed by regressing the anomaly at each pressure by the tropical mean surface temperature anomaly. The dashed lines represent the $2\sigma$ range on the regression coefficients computed using a bootstrapping method as described in the text.
Figure 6. Sensitivity of tropical mean (a) temperature, (b) water vapor mixing ratio, (c) radiative cooling, (d) static stability, (e) diabatic subsidence, and (f) diabatic convergence to tropical mean surface temperature. Sensitivity profiles are computed by regressing the anomaly at each pressure by the tropical mean surface temperature anomaly. The dashed lines represent the 2σ range on the regression coefficients computed using a bootstrapping method as described in the text.
Figure 7. (blue) Sensitivity of tropical mean (a) cloud top frequency of occurrence from CloudSat and cloud fraction from (b) MODIS, (c) AIRS, and (d) ISCCP to tropical mean surface temperature. Only clouds with \( \tau \geq 1.3 \) are included in the ISCCP cloud fraction plot. Overlaid in red in each panel is the sensitivity of diabatic convergence to tropical mean surface temperature as shown in Figure 6f. Note that the range of values on the upper x-axis varies from panel to panel. The dashed lines represent the \( 2\sigma \) range on the regression coefficients computed using a bootstrapping method as described in the text.
Figure 8. (blue) Tropical mean (a) pressure, (b) water vapor mixing ratio, (c) radiative cooling, (d) static stability, (e) diabatic subsidence, and (f) diabatic convergence, along with (red) the sum of the mean profiles and the perturbation profiles shown in Figure 6, all plotted as functions of tropical mean temperature from AIRS-MLS. Dashed red lines represent the 2σ range of uncertainty on the perturbation profile.
Figure 9.  (a) Tropical mean (blue) CloudSat cloud top frequency of occurrence and (red) diabatic convergence. The dashed lines represent the mean profile and the solid lines represent the sum of the mean and perturbation profile shown in panel b. (b) Sensitivity of tropical mean (blue) CloudSat cloud top frequency of occurrence and (red) diabatic convergence to tropical mean surface temperature. The dashed lines represent the $2\sigma$ range on the regression coefficients computed using a bootstrapping method as described in the text.
Figure 10. (a) Sensitivity of tropical mean MODIS cloud fraction in each cloud top pressure and visible optical depth bin to tropical mean surface temperature. Product of the cloud fraction sensitivity shown in (a) with overcast sky (b) LW, (c) SW, and (d) net cloud forcing histograms. The sensitivities of each quantity to tropical mean surface temperature computed by summing the histograms are shown in the titles. The contributions of high (CTP < 450) cloud anomalies are given in parenthesis.
Table 1. Slopes of regression lines derived between $T_{sfc}$ anomalies and cloud radiative forcing anomalies from several satellite datasets, along with cloud feedback estimates derived from the anomalous MODIS cloud fraction histogram (in W m$^{-2}$ K$^{-1}$). The errorbars represent the 2$\sigma$ ranges on the regression slopes computed using a bootstrapping method as described in the text. Note that the histogram-derived values are estimates of the response of TOA fluxes to clouds alone whereas cloud radiative forcing changes measured by CERES and AIRS include a small impact of non-cloud-induced flux changes.

<table>
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<tr>
<th>Dataset</th>
<th>Temporal Coverage</th>
<th>$LWCF$ vs. $T_{sfc}$</th>
<th>$SWCF$ vs. $T_{sfc}$</th>
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<tbody>
<tr>
<td>CERES SYN “lite” (Terra)</td>
<td>Mar 2000 - Feb 2010</td>
<td>-1.4 ± 0.6</td>
<td>1.9 ± 1.0</td>
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<td>CERES SSF “lite” (Terra)</td>
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<td>CERES SRBAVG (Aqua)</td>
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<tr>
<td>MODIS (Aqua) + Fu-Liou (High Only)</td>
<td>Sep 2002 - Jul 2010</td>
<td>-1.0 ± 1.3</td>
<td>1.4 ± 1.1</td>
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