Total Heating Characteristics of the ISCCP Tropical and Subtropical Cloud Regimes

JUSTIN P. STACHNIK and COURTNEY SCHUMACHER
Texas A&M University, College Station, Texas
PAUL E. CIESIELSKI
Colorado State University, Fort Collins, Colorado

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ABSTRACT

Composite profiles of the apparent heat source $Q_1$ and moisture sink $Q_2$ are calculated for the International Satellite Cloud Climatology Project (ISCCP) cloud regimes (or “weather states”) using sounding observations from 10 field campaigns comprising both tropical and subtropical domains. Distinct heating profiles were determined for each ISCCP cloud regime, ranging from strong, upper-tropospheric heating for mesoscale convective systems (WS1) to integrated cooling for populations typically associated with marine stratus and stratocumulus clouds (WS5, WS6, and WS7). Despite being primarily associated with thin cirrus, the corresponding regime (WS4) has heating maxima in the lower and midtroposphere due to the presence of underlying clouds. Regime-averaged $Q_2$ profiles showed similar transitions with strong drying observed for deep convection and low-level moistening for marine boundary layer clouds. The derived profiles were generally similar over land and ocean with the notable exception of the fair-weather cumulus regime (WS8). Additional midlevel moistening was identified for several weather states over land, suggesting enhanced detrainment and more frequent congestus clouds compared to oceanic domains.

A control simulation using the Community Atmosphere Model, version 4 (CAM4), was similar to the large-scale patterns of diabatic heating at low levels produced by the ISCCP composites. Differences were more pronounced at middle and upper levels and are largely attributed to the uncertainty in the heating profiles for the cumulus regime (WS8). Low-level heating anomalies were calculated for each phase of the Madden–Julian oscillation (MJO) and they precede upper-tropospheric heating from deep convection by 3–4 phases. Implications for future research using ISCCP heating reconstructions are also discussed.

1. Introduction

It is well known that clouds play an important role in controlling the daily weather, yet the aggregate effects and associated climate feedbacks of cloud systems remain less understood. These feedbacks are especially relevant in the tropics, where the total diabatic heating produced by clouds and precipitating systems directly couples these phenomena to the large-scale circulation. Variations in the magnitude and spatial distribution of heating from tropical cloud clusters elicit a different dynamical response in numerical models at both regional and global scales (e.g., Hartmann et al. 1984; Lin et al. 2004; Schumacher et al. 2004; Lappen and Schumacher 2012). Determining an accurate horizontal and vertical distribution of tropical heating is therefore paramount to better understanding and predicting climate variability in general circulation models (GCMs).

A large number of studies have focused on the calculation of apparent diabatic heating from cloud systems (comprising latent heating associated with phase changes of water, radiative processes, and eddy sensible heat fluxes) using data from intensive observation periods in tropical field campaigns (e.g., Yanai et al. 1973; Johnson 1976; Thompson et al. 1979; Lin and Johnson 1996; Zhang et al. 2001; Johnson et al. 2010). These studies typically rely on network measurements of temperature and wind across a large-scale domain (on the order of 100 000 km$^2$) and are inherently restricted to the cloud characteristics of a specific region during periods of active sampling. Datasets including gridded model output (e.g., Nigam et al. 2000) and reanalyses (Sardeshmukh 1993), which are strongly influenced by
cumulus parameterizations—a source of model infidelity, can be used for the above calculations, though the derivative profiles often contain significant variability among datasets and large heating differences compared to observations (e.g., Chan and Nigam 2009; Jiang et al. 2011). Other studies using direct measurements of cloud systems including precipitation radars (e.g., Houze 1982, 1989) and special ground-based radiation measurements (e.g., Li et al. 2013) are also limited to single points and may not be representative of the general cloud population across a larger domain.

Recent improvements in diabatic heating estimates from the Tropical Rainfall Measuring Mission (TRMM) satellite show significant promise for real-time global monitoring (Tao et al. 2006; L’Ecuyer and McGarragh 2010). A number of algorithms use observations from the TRMM satellite (e.g., Tao et al. 2001; L’Ecuyer and Stephens 2003; Shige et al. 2004; Olson et al. 2006) along with reference profiles mostly derived from cloud-resolving models to estimate grid-averaged latent and radiative heating profiles across the tropics. These techniques, however, face certain limitations. First, they rely on model output as the basis of their lookup tables so include any model errors in their estimates. In addition, the retrievals struggle with areas of weak intensity clouds (including shallow convection and stratiform cloud) because of the footprint resolution and sensitivity of the satellite instruments. In particular, the redistri-

bution of low-level latent heating from nonprecipitating cumulus clouds and upper-level radiative heating associated with anvil cloud are both necessary components of the total heating in order to achieve a more realistic large-scale response in GCMs (e.g., Schumacher et al. 2004). TRMM heating has also been observed to be too weak compared to observations (Chan and Nigam 2009), although improvements have been made in this regard (Tao et al. 2010). Finally, the TRMM retrievals are limited in space and time resolution by the sampling of the satellite.

This study presents a method based solely on observations for determining the four-dimensional total diabatic heating field at up to 3-h resolution by compositing profiles from numerous tropical and subtropical field campaigns and matching to cloud regimes or “weather states” (Jakob and Tselioudis 2003; Jakob et al. 2005; Rossow et al. 2005) from a 25-yr subset of the International Satellite Cloud Climatology Project (ISCCP; Schiffer and Rossow 1983). The ISCCP weather states represent a statistical set of physically identifiable and recurring cloud mixtures over a large area [∼(280 km × 280 km)] with populations ranging from large mesoscale convective systems (MCSs) to nonprecipitating, boundary layer cumulus. Previous studies have examined the latent and radiative heating characteristics of the ISCCP cloud regimes (Jakob et al. 2005; Jakob and Schumacher 2008; Oreopoulos and Rossow 2011; Li et al. 2013); this study is the first to investigate the total diabatic heating associated with each weather state. Compositing the diabatic heating profiles by ISCCP regime provides the added benefit of determining which mixture of cloud types has the greatest impact upon tropical variations in diabatic heating (i.e., the ability to identify what phenomena comprise the mean heating), a subject that has received significantly less attention in the literature (e.g., Schumacher et al. 2008). Furthermore, many GCMs now include an ISCCP simulator (Klein and Jakob 1999) and the ability to diagnose unique heating profiles separated by cloud type provides a new observational metric for the verification of modeled clouds as sorted by regime (e.g., Webb et al. 2001; Williams and Webb 2009).

This study advances the idea proposed by Jakob et al. (2005) to extend point measurements of the ISCCP weather states to other regions of the globe based on average cloud properties and their relative frequency of occurrence. As such, we first determine average heating rates for the tropical and subtropical cloud regimes based on a large database of nearly 3000 sounding budget profiles from field campaigns and then evaluate the feasibility of reconstructing global heating patterns using this data. Section 2 describes the data and details of the compositing used in this study. The resulting heating profiles are presented in section 3, with a strong emphasis on the ensemble average in lieu of discussing individual field campaigns. Section 4 demonstrates two potential applications for studying tropical and subtropical clouds and precipitation using the heating reconstructions. Section 5 closes with a summary of key results and identifies avenues for continued work.

2. Data and methods

a. ISCCP weather states

The ISCCP D1 dataset (Rossow and Schiffer 1999) contains a global climatology of cloud properties (including cloud amount, cloud-top pressure, and cloud optical thickness, among others) derived from visible and infrared radiances observed by geostationary and polar-orbiting satellites. Jakob and Tselioudis (2003) and Rossow et al. (2005) supported the idea that the appearance of commonly recurring cloud regimes (also referred to as weather states) could be used as a proxy for multivariate dynamical states of the tropical atmosphere. Studies since continue to link the ISCCP weather states to dynamical regimes and synoptic weather
phenomena (e.g., Gordon and Norris 2010; Mekonnen and Rossow 2011) in addition to documenting longer-term variability of convectively active and suppressed regimes in the tropics (Tselioudis and Rossow 2011).

In short, the ISCCP weather states use a $k$-means clustering algorithm (Anderberg 1973) to identify repeating patterns of cloud height and extinction co-variations over a large area [$\sim(280\,\text{km} \times 280\,\text{km})$] from individual satellite pixels of about 5 km in size. Joint histograms of cloud-top pressure and optical thickness are produced every three hours at each $2.5^\circ \times 2.5^\circ$ grid point from July 1983 to June 2008 and sorted into a predefined number of groups (details provided in Rossow et al. 2005).

The subtropical extension of the ISCCP regimes identifies eight distinctive weather states (WS1–WS8) for all longitudes spanning the $35^\circ\text{N}$–$35^\circ\text{S}$ domain. The average centroids (i.e., mean histogram) for each cluster are shown in Fig. 1. Shading represents the average frequency of occurrence for clouds falling into specific height and extinction bins within each regime. The integral across all bins or cluster cloud fraction (CCF) identifies the average total cloud cover for each cluster. Grid points may also be identified as a separate weather state if the entire field of view is clear (WS0) or data are flagged as missing (WS98). The corresponding relative frequency of occurrence (RFO) for each regime (i.e., the number of counts for a particular weather state divided by the total number of counts for nonmissing, daytime data) is listed at the top of each panel. Figure 2 shows the spatial distribution of the annual average RFO, along with markers identifying the center of the field campaign heating domains described in the following subsection, for each weather state.

Based on the average cloud properties and geographic distribution in Figs. 1 and 2, our interpretation of the weather states is as follows: WS1 comprises a population of tall, optically thick clouds primarily concentrated along the intertropical convergence zone (ITCZ). This weather state describes vigorous, deep convection with extensive cirrostratus and stratiform precipitation (near 100% CCF) normally associated with tropical MCSs. WS2 generally occurs in the same regions as WS1 and has a blend of tall, convective clouds with cirrus anvils of moderate optical thickness (cumulonimbus/cirrostratus). A mix of less vigorous convection and midlevel clouds (cumulonimbus/congestus) is apparent in WS3, though its RFO (17.6%) is significantly higher than WS1 and WS2 (5.9% and 8.3%, respectively). WS4 occurs over land and ocean domains, primarily containing thin cirrus with tall heights and low optical thickness. WS5 and WS6 represent moderately thick, low clouds principally found over the ocean, characteristic of marine boundary layer stratocumulus and stratus, respectively. These regimes have low RFOs (6.7% and 4.5%) and are typically restricted to areas of cold sea surface temperatures (SSTs) west of the continents. WS7 represents a transition state with low cloud tops and low to moderate optical thickness typical of broken stratocumulus and cumulus clouds (CCF of 59.7% compared to 84.7% and 74.5% for WS5 and WS6, respectively). WS7 occurs mainly over the oceans, yet has a stronger coastal influence than the previous two regimes. Finally, WS8 contains low cloud tops and small values of optical thickness (i.e., populations of mostly nonprecipitating cumulus) and has the largest domain average RFO of all the weather states (35.0%). WS8 also has significant cloud-free area (CCF of 24.1%) due to the expansive-ness of this regime and its convective organization (e.g., cloud streets, open and closed cells).

It is worth repeating that the ISCCP weather states represent mixtures of clouds within the larger $2.5^\circ \times 2.5^\circ$ area, hence the distribution of cloud properties shown in Fig. 1. While the preceding classifications describe the predominant cloud type within the grid domain, there is likely overlap in the distributions between weather states (e.g., WS4 will contain low-level clouds and WS8 will include some cumulus congestus). Nevertheless, previous studies have shown that the ISCCP weather states have unique thermodynamic and precipitation characteristics (Jakob et al. 2005; Lee et al. 2013) in addition to different latent heating and radiative properties (Jakob and Schumacher 2008; Oreopoulos and Rossow 2011; Li et al. 2013). The total diabatic heating profiles should likewise vary significantly among regimes.

Finally, one caveat of the ISCCP weather states is that the classifications are only available during daytime hours when passive satellite instruments can determine visible optical thickness. A major goal of this work is the ability to reconstruct global diabatic heating fields for future studies of tropical climate dynamics (section 4), so we extend the first and last regime classifications for every day and grid point backward and forward 6 h, respectively. Jakob et al. (2005) showed that the tropical weather states have persistence on the order of a day. Furthermore, we note that the average 6-h persistence for individual regimes during daytime hours using the dates and locations of the field campaigns in this study is 47.6%. Persistence rates significantly increase when grouping weather states considered convectively active or suppressed (e.g., Tselioudis and Rossow 2011), and we find many instances in our data where the same regime is identified for a given field campaign uninterrupted for multiple days. Although this assumption may blur some of the results, the shapes of the
individual composites tend to be mostly similar to those calculated using daytime-only data. As will be shown, the derived heating profiles have distinct differences among weather states, suggesting the persistence assumption remains effective for determining the relative heating contribution from each cloud regime.

b. Large-scale budget data

The presence of diabatic processes results in changes to the dry static energy $s$, which is equal to the sum of the enthalpy and potential energy ($s = c_p T + gz$). The apparent heat source $Q_1$ represents the net heating due to
the ensemble of convection within a region (e.g., a sounding network) including a correlation term resulting from unresolved eddies. Yanai et al. (1973) defined the apparent heat source as follows:

\[ Q_1 = \frac{\partial s}{\partial t} + \nabla \cdot (s \mathbf{v}) + \frac{\partial (s \omega)}{\partial p} = Q_R + L(c - e) - \frac{\partial (s \omega)}{\partial p}, \]

where \( Q_R \) is the radiative heating or cooling or the atmosphere, \( L(c - e) \) is the contribution of latent heating or cooling from associated phase changes of water, and \( s \omega \) is the vertical transport of sensible heat by small-scale eddies. Overbar quantities represent horizontal means across the averaging area and primes indicate small-scale deviations. A similar equation can be written following moisture conservation. Yanai et al. (1973) define the apparent moisture sink \( Q_2 \) as

\[ Q_2 = -L \left[ \frac{\partial q}{\partial t} + \nabla \cdot (q \mathbf{v}) + \frac{\partial (q \omega)}{\partial p} \right] = L(c - e) + L \frac{\partial (q \omega)}{\partial p}, \]

where \( q \) is the specific humidity.

Calculations of \( Q_1 \) and \( Q_2 \) can be made using Eqs. (1) and (3) as the residual quantities of sounding network measurements of temperature, wind, and moisture (Yanai et al. 1973; Johnson 1976; Thompson et al. 1979; among others). Traditional budget studies are adversely affected by instrumental errors in radiosonde measurements and random sampling errors that alias small-scale variations in the winds and state variables on to larger scales (e.g., Mapes et al. 2003). In an effort to minimize these errors, Zhang and Lin (1997) developed a variational analysis technique that adjusts sounding observations within the range of measurement and instrument uncertainties to satisfy column-integrated budgets of mass, energy, and moisture. Though only the smallest possible adjustments are made, significant differences may appear in the resultant \( Q_1 \) and \( Q_2 \) profiles. More information on the variational analysis technique and comparison to traditional budget method is reviewed in Zhang et al. (2001).

Profiles of the apparent heat source and moisture sink from field experiments overlapping the 25 yr of weather state data are used in this study. Campaigns (listed in chronological order) include the Atlantic Stratocumulus Transition Experiment (ASTEX), Tropical Ocean Global Atmosphere Coupled Ocean–Atmosphere Response Experiment (TOGA COARE), South China Sea Monsoon Experiment (SCSMEX), TRMM Large-Scale Biosphere–Atmosphere Experiment (LBA), Kwajalein

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**Fig. 2.** Geographic distribution of the annual average relative frequency of occurrence for each weather state from 1984 to 2007. Markers (circled times symbol) indicate the ISCCP grid point nearest the center of the budget domain for the field campaigns identified in Table 1.
Experiment (KWAJEX), North American Monsoon Experiment (NAME), Tropical Warm Pool International Cloud Experiment (TWP-ICE), African Monsoon Multidisciplinary Analysis (AMMA), Mirai Indian Ocean Cruise for the Study of the MJO Onset (MISMO), and Terrain-Induced Monsoon Rainfall Experiment (TIMREX). The average campaign $Q_1$ profiles (including an equally weighted ensemble average) are provided for context in Fig. 3 and demonstrate the innate variability among domains prior to subsetting by weather state. Data availability, domain details, and references for the initial budget calculations (including discussion of the campaign average profiles) are summarized in Table 1.

Data from ASTEX, TOGA, SCSMEX, NAME, AMMA, MISMO, and TIMREX use traditional budget techniques relying solely on observations. Several of the projects have subdomains with profiles available for geographically different regions. These include separate basins over the northern and southern South China Sea (SCSMEX-N and SCSMEX-S), northern and southern regions of inland and coastal West Africa (AMMA-N and AMMA-S), and land and ocean areas surrounding

![Figure 3](image)

**Fig. 3.** Campaign average total diabatic heating ($Q_1$) profiles for the domains used in this study. Profiles for individual campaigns over mostly-land and mostly-ocean domains are indicated using solid and dashed lines, respectively. The thick line represents an equally weighted ensemble average for all domains.

<table>
<thead>
<tr>
<th>Project</th>
<th>Data availability</th>
<th>Samples (No.)</th>
<th>Domain center</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>ASTEX</td>
<td>1 Jun–15 Jun 1992</td>
<td>161 (3 h)</td>
<td>33.75°N, 22.25°W</td>
<td>Ciesielski et al. (1999)</td>
</tr>
<tr>
<td>KWAJEX</td>
<td>24 Jul–14 Sep 1999</td>
<td>210</td>
<td>8.75°N, 166.25°E</td>
<td>Schumacher et al. (2007)</td>
</tr>
<tr>
<td>TWP-ICE</td>
<td>17 Jan–12 Feb 2006</td>
<td>210 (3 h)</td>
<td>11.25°S, 131.25°E</td>
<td>Xie et al. (2010)</td>
</tr>
<tr>
<td>AMMA</td>
<td>1 Jun–30 Sep 2006</td>
<td>484</td>
<td>13.75°/6.75°N, 3.75°/3.75°E</td>
<td>Xiping et al. (2013)</td>
</tr>
<tr>
<td>TIMREX</td>
<td>15 May–26 Jun 2008</td>
<td>169</td>
<td>23.75°N, 118.75°E</td>
<td>Ruppert et al. (2013)</td>
</tr>
</tbody>
</table>
Taiwan (TIMREX-L and TIMREX-O). Variational analysis was used to determine the profiles from TRMM-LBA, KWAJEX, and TWP-ICE. Additional budget calculations using variational analysis were available for TOGA and the SCSMEX-N domains. The corresponding $Q_1$ and $Q_2$ profiles for these regions were averaged with the estimates using the traditional budget approach for matching times before compositing by ISCCP regime. Although there were notable differences in the magnitude of heating between the traditional budgets and variational datasets (largely attributed to differences in the domain size), the shapes of the profiles were similar and both datasets were used in an effort to create a more realistic consensus.

Figure 2 shows the centers of the budget domains for each of the experiments listed in Table 1, along with the annual average RFO for each of the ISCCP weather states. Many of the project domains are well placed to study the full range of ISCCP regimes, though reliable data for WS5, WS6, and WS7 are limited to the ASTEX region (i.e., open ocean areas northwest of Africa). While the ISCCP cloud properties may vary by region and show some seasonality, the budget studies range across a wide array of tropical and subtropical locations and include sampling from all times of the year. In addition, the compositing technique should eliminate any regional or seasonal bias, thereby producing a profile representative of the annual average, mean global cloud regime.

Special processing was necessary for the ASTEX dataset, which only contained $Q_1$ and $Q_2$ estimates from the surface to 4 km. As the ASTEX region is dominated by shallow, marine boundary layer clouds, the total diabatic heating should be well represented by the radiative component in middle and upper levels. Consequently, we supplement the original $Q_1$ data above 4 km with the climatological values of $Q_R$ over the ASTEX region during June using TRMM data from 1998–2010 following L’Ecuyer and Stephens (2003) and L’Ecuyer and McGarragh (2010).

All profiles were linearly interpolated to a constant pressure grid ranging from 1000 to 100 hPa in 25-hPa increments. Data at pressure levels below the surface were ignored. Most datasets include $Q_1$ and $Q_2$ calculations four times daily (at 0000, 0600, 1200, and 1800 UTC) and only those times were used when compositing with the ISCCP data. The $Q_1$ and $Q_2$ profiles were available every 3 h for ASTEX and TWP-ICE with the compositing technique using all available profiles from these projects. Campaigns with less than 16 occurrences of a particular regime (including nighttime hours assigned their daytime adjacent classification) were disregarded based on subjective estimates of profile noise and overall representativeness. The numbers of samples for each experiment and ISCCP classification are provided in Table 2. Although the budget domains occasionally span large regions with multiple ISCCP grid points, only the classifications nearest the center of the domain are used in the compositing technique.

No additional averaging was performed for the individual field campaigns. Equally weighted ensemble average profiles were calculated for each regime over
mostly-land (TRMM-LBA, NAME, TWP-ICE, AMMA-N, AMMA-S, and TIMREX-L), mostly-ocean (ASTEX, TOGA, SCSMEX-N, SCSMEX-S, KWAJEX, MISMO, and TIMREX-O), and all domains. Additional experiments calculating the ensemble mean with a weighted average (e.g., by number of observations) produced similar results as the equally weighted technique when using a minimum threshold filter.

c. Other data

A 24-yr control simulation was performed as a reference for global diabatic heating using the Community Atmosphere Model, version 4 (CAM4). The model framework is identical to that utilized by Lappen and Schumacher (2012), with a time step of 1800 s, 26 vertical levels, and horizontal resolution of 1.9° latitude × 2.5° longitude using prescribed SSTs. A modified Zhang and McFarlane (1995) convective parameterization was used, in addition to boundary layer physics from Holtslag and Boville (1993) and a shallow convection scheme following Hack (1994). Monthly output of the total diabatic heating (and individual components) were produced from 1984 to 2007 and regridded to match the 2.5° × 2.5° ISCCP domain using an area-conservative remapping function.

Finally, the Wheeler and Hendon (2004) index is used to monitor the magnitude and phase of the Madden–Julian oscillation (MJO; Madden and Julian 1994) during November–April of 1983–2008. Strong MJO events were identified as those days with a total real-time multivariate MJO (RMM) amplitude greater than or equal to one. Heating anomalies for strong events were calculated by MJO phase for each of the ISCCP regimes relative to the climatological background heating. Results from the modeling and MJO components of this study are presented in section 4.

3. Results

a. $Q_1$ profiles

Composite profiles of the apparent heat source for each project meeting the minimum sample threshold and the ensemble average for all domains are shown in Fig. 4. It is clear that WS1 (MCSs) has the largest ensemble average heating rate of all the regimes. Maximum heating occurs near 450 hPa and peaks near 8 K day$^{-1}$, more than twice the heating maxima from any other regime. The result is not surprising, however, as latent heating from precipitation production is the dominant component of the total diabatic field. WS1, which has the greatest average precipitation rate ($\sim$19 mm day$^{-1}$; Lee et al. 2013), would thus be expected to have the strongest latent component and total heating for the ISCCP regimes. The profile shape is top heavy and likely explained by the presence of a significantly high stratiform rain fraction. Houze (1982, 1989) showed that higher stratiform rain fractions lead to stronger heating in the upper levels of the atmosphere because of particle growth by deposition, with regions of cooling (and melting) below. Moreover, the result is in agreement with Jakob and Schumacher (2008), who found the analogous WS1 for a separate weather state study over the tropical west Pacific had stratiform rain fractions ranging from 50% to 70%. In addition, there is significant anvil radiative heating for WS1 (Li et al. 2013) contributing to the overall profile shape. The presence of deep heating throughout the entire troposphere (Fig. 4a) also suggests that areas of convective rain (with heating maxima in the midtroposphere) are present alongside the stratiform rain area.

Although WS1 comprises the widest distribution of shortwave, longwave, and net cloud radiative forcing when compared to the other weather states (Oreopoulos and Ros sow 2011), the substantial variability among field campaigns is best explained by the differences in rain rate. For example, a single nocturnal event with extreme rain rates ($>8$ mm h$^{-1}$) during TWP-ICE resulted in the anomalously large average value for this campaign and the mean profile becomes similar to the ensemble average when excluding this event.

The profile for WS2 has a similar shape to WS1, though peak heating for the ensemble average only reaches 3 K day$^{-1}$ (note the scale difference for Fig. 4a). This is in agreement with its significantly smaller average rain rate ($\sim$5 mm day$^{-1}$; Lee et al. 2013) and fewer clouds with large optical thickness (cf. Figs. 1a,b). The peak heating occurs at 400 hPa, slightly higher than WS1. This difference may be explained by a higher percent stratiform rain and nonprecipitating anvil, as a weakening system initially classified as WS1 would produce an extensive stratiform area as it transitions to WS2. Rickenbach (2004) documented that MCSs during TRMM-LBA usually formed along the coast of Brazil and moved westward into the sampling domain well after reaching their convective peak. The extreme stratiform-like profile for TRMM-LBA in Fig. 4b exemplifies this point, and the number of samples for WS2 in TRMM-LBA is 3 times as many for WS1 (Table 2). WS3 also achieves a peak heating of 3 K day$^{-1}$, but has a broad maximum from 600 to 400 hPa. The lowering height of the heating maximum is best explained by the mixed and weaker nature of WS3 convection (cumulonimbus/congestus) and a smaller stratiform rain fraction (Jakob and Schumacher 2008). There is also considerable variability among campaigns due in part to differing rain rates.
WS4 represents thin cirrus and has weak heating (less than 1.5 K day$^{-1}$) for the ensemble average throughout the entire troposphere. Although many of the individual domains have small heating peaks above 200 hPa, the primary heating peaks are in the lower (950 hPa) and middle (500–400 hPa) troposphere. The near-surface peak is likely attributed to sensible heat fluxes over land domains and the redistribution of this heat by turbulent eddies and shallow cumulus that often accompany tropical cirrus (e.g., Jakob et al. 2005; Schumacher et al. 2008). The midlevel peak may be a result of thin anvil and outflow from tropical convection reaching only moderate heights, consistent with the cloud type heating profiles of Schumacher et al. (2008) derived from visual observations during KWAJEX.

WS5, WS6, and WS7 are responsible for significant cooling throughout most of the atmosphere. WS5 and WS6 have similar profile shapes with weak near-surface heating (less than 1 K day$^{-1}$) up to 950 hPa, with cooling of 5 and 3 K day$^{-1}$ at 850 hPa, respectively. WS7 has
slightly stronger surface heating to 900 hPa and a similar cooling peak (3.5 K day\(^{-1}\) at 825 hPa), resulting in a smaller net cooling than the previous two regimes. These results are consistent with the budget calculations and net surface and top of atmosphere longwave cloud radiative effects determined for the ISCCP extended low-latitude weather states by Oreopoulos and Rossow (2011). Although the composites are limited to the ASTEX region,\(^1\) the shapes and magnitude of the \(Q_1\) profiles agree with prior work. For example, Nitta and Esbensen (1974) observed shallow, nonprecipitating stratocumulus clouds in the eastern Caribbean during the Barbados Oceanographic and Meteorological Experiment (BOMEX), with results similar to ours. The remainder of the profile above 4 km (~650 hPa) is the climatological \(Q_R\) derived from TRMM data (see section 2), which resembles typical clear-sky radiative cooling with values generally around 1.25 K day\(^{-1}\).

The average profile for WS8 shows a maximum of 1 K day\(^{-1}\) at 950 hPa with very weak heating observed throughout the remainder of the troposphere. Although these cloud types are generally nonprecipitating and have zero net latent heating, the convective mass flux is capable of redistributing heating, with low-level warming from eddy transport and condensation in cloud, and cooling above due to evaporation and detrainment (e.g., Nitta and Esbensen 1974). There is significant spread among the individual heating estimates in the middle levels (up to 7 K day\(^{-1}\)), and this uncertainty is discussed in more detail below.

The fidelity of these results was further examined by calculating the ISCCP regime average profiles excluding the \(Q_1\) dataset from a particular field campaign and reconstructing the observed mean heating profile for that field campaign as a function of weather state frequency of occurrence. The predicted reconstructions and sounding-based campaign average profiles (excluding ASTEX) are shown in Fig. 5. Generally, most of the predictions have zero net latent heating, the convective mass flux is capable of redistributing heating, with low-level warming from eddy transport and condensation in cloud, and cooling above due to evaporation and detrainment (e.g., Nitta and Esbensen 1974). There is significant spread among the individual heating estimates in the middle levels (up to 7 K day\(^{-1}\)), and this uncertainty is discussed in more detail below.

Differences in the mean profile for regimes 1–4 and 8 over mostly-land and mostly-ocean domains are shown in Fig. 6. All of the regimes have a tendency for enhanced heating at low levels (~1 K day\(^{-1}\)) over mostly-land domains. Stronger sensible heating and an enhanced diurnal cycle over land likely cause this result. The shape and magnitude of the mean heating profile for WS1 is nearly identical over land and ocean (Fig. 6a). This result is consistent with previous work, as Lee et al. (2013) documented comparable precipitation rates over both domains when excluding nonprecipitating pixels (i.e., zero rain rate) from their averaging.

The mean profiles for WS2 show enhanced heating over the ocean from 875 to 550 hPa, with less heating relative to land domains from 550 to 100 hPa (Fig. 6b). These differences are likely nonphysical, as the WS2 land composite is strongly influenced by the individual heating profile from TRMM-LBA. As previously mentioned, the TRMM-LBA retrieval is biased toward extreme stratiform rain fractions given the characteristic evolution of South American MCSs developing along the Brazilian coast and moving westward into the analysis domain well after reaching their convective peak.

\(^{1}\)Sufficient numbers of samples above the minimum threshold were available from AMMA-S for WS5 and AMMA-N for WS6 and WS7. These profiles were omitted from the analysis, however, as they exhibited strong near-surface effects and contained heating of several kelvins per day throughout the entire troposphere. These profiles were deemed unrepresentative of the main WS5, WS6, and WS7 populations, which primarily occur over the oceans (Fig. 2). Similar data were also rejected for WS7 from MISMO that more closely resembled the heating profiles for altocumulus clouds determined by Schumacher et al. (2008).
The profile shapes are similar for WS3 (Fig. 6c), with the exception of stronger near-surface heating identified over land in AMMA-N and AMMA-S (Fig. 4c). The ocean profile contains slightly more heating in middle levels relative to land (≈1 K day$^{-1}$) and is likely due to the presumed heavier rain rates for MISMO and TIMREX-O (Fig. 4c). The land and ocean composites for WS4 (Fig. 6d) also have similar shapes to the ensemble average, though the heating maxima are 1.5–2.5 K day$^{-1}$ greater over land at both the low- and midlevel peaks. The enhanced sensible heat flux over the AMMA-N and AMMA-S domains likely accounts for the lower peak, though an appropriate reason for differences in the middle levels is not immediately clear. It is possible that there is a greater frequency of more dense (i.e., higher optical thickness) cirrus anvils and outflow from convection peaking near 500 hPa over land, though this hypothesis requires further investigation.

The domain differences in $Q_1$ are most pronounced for WS8 (Fig. 6e). There is weak heating ($<1$ K day$^{-1}$) over the ocean from the surface to 925 hPa, with weak cooling throughout the rest of the troposphere. This profile matches the expected shape for nonprecipitating, shallow cumulus at low levels with clear-sky radiative cooling above. The land composite meanwhile has peak heating of $\sim$2.5 K day$^{-1}$ between 950 and 900 hPa with moderate warming (1–2 K day$^{-1}$) up to 200 hPa. A broad heating maximum exists from 600 to 400 hPa for the land-only composite, similar to the average profile for WS3 (i.e., cumulonimbus/congestus). Furthermore, all of the individual land domains show heating throughout the entire troposphere, while ocean areas primarily contain cooling above 925 hPa (Fig. 4h). Oreopoulos and Rossow (2011) identified WS8 as having the smallest mean cloud radiative effect compared to all other regimes, indicating a weak cooling of the atmosphere. The above finding is consistent with our integrated total diabatic heating for ocean domains.

Lee et al. (2013) found the average rain rate for WS8 was greater over land than ocean. Although the mean
rain rates for both domains were small (<1 mm day\(^{-1}\)), these totals included areas without any precipitation (i.e., zero rain rates) and 80%–90% of their WS8 pixels were nonprecipitating. Those areas where WS8 produced heavier precipitation (coincident with our land domains) had average rain rates of 2–5 mm day\(^{-1}\), comparable to those from WS3 over the same regions.

Given the predominance and persistence of WS8 over the oceans (Fig. 2h), along with clustering at low heights and weak-to-moderate optical thickness (Fig. 1h), we expect the cloud field over the ocean domain would be mostly homogenous. A similar “most likely” probability exists for clouds of low height and low thickness over land, but prominent changes in surface use or other heterogeneities could allow certain pixels within the larger population to grow to higher heights or achieve larger values of optical thickness (i.e., the “tails” of the cluster distribution in Fig. 1h). The joint histograms for these regions might look similar to WS3, excluding the cumulonimbus population in the upper right (Fig. 1c), and their overall distribution and lower cloud fraction would result in their being categorized as WS8. The previous scenario, in tandem with the observed increase in precipitation rates, could potentially explain the differences between the \(Q_1\) profiles for our land and ocean domains. Nevertheless, verification of the above hypothesis is beyond the scope of this work.

b. \(Q_2\) profiles

Composites of the apparent moisture sink for each regime are shown in Fig. 7. Estimates of \(Q_2\) for individual campaigns generally demonstrate more variability than \(Q_1\) (Lin and Johnson 1996; Johnson and Ciesielski 2000), and the following discussion focuses on relative variability between regimes.

There is strong drying (positive values) throughout the entire depth of the troposphere for WS1 (Fig. 7a). The profile has two peaks with a maximum of 3.5 and 5 K day\(^{-1}\) at 825 and 475 hPa, respectively. As identified earlier, there is significant convective and stratiform rain associated with WS1. The dual peak is consistent with Johnson (1984), where low-level drying occurs in convective updrafts and drying at upper levels is a result of mesoscale lifting within the stratiform region. As with
The shapes of the drying profile for WS2 and WS3 are similar to WS1, though the peak amplitudes are greatly reduced (1.75 and 2.5 K day\(^{-1}\) for each regime, respectively). WS2 has a top-heavy drying profile due to the increased stratiform rain fraction, while the WS3 profile is more bottom heavy as the convection becomes weaker and contains a lower fraction of stratiform rain (similar to the trends for \(Q_1\)). The ensemble average profile for WS4 is near zero throughout the entire troposphere.

There is primarily moistening (negative values) throughout the lower atmosphere for WS5, WS6, and WS7 (Figs. 7c–g) attributed to surface evaporation and upward transport of water vapor by clouds. Maritime stratus and stratocumulus are persistent across local regions (Figs. 2e,f) and have average precipitation rates less than 1 mm day\(^{-1}\) (Lee et al. 2013). Variations in the magnitude of the low-level moistening among WS5–WS7 reflect differences in the balance of evaporation (adding moisture) and precipitation (removing moisture) in these regimes. WS5 has a slightly higher average precipitation rate (Lee et al. 2013) and displays a weaker peak moistening of 4 K day\(^{-1}\) compared to the 6 K day\(^{-1}\)
maximum for WS6 and WS7. The ensemble average drying profile for WS8 is particularly noisy with weak moistening and drying (<1 K day\(^{-1}\)) below and above 600 hPa, respectively.

Figure 8 shows the equivalent average drying profiles over land and ocean domains. Low-level moistening is apparent for WS2, WS3, WS4, and WS8 over the oceans because of increased eddy transport of near-surface water vapor. The magnitude and height of the moistening layer vary by regime and are most pronounced for WS8 (Fig. 8e). Moistening over oceans exceeds 2 K day\(^{-1}\) at 875 hPa for the cumulus regime, generally consistent with Schumacher et al. (2008). WS8 has relatively large drying (~3 K day\(^{-1}\)) throughout the low levels over land, consistent with the idea that these regimes precipitate more easily and have larger average rain rates over these domains.

Another common feature of the land and ocean profiles is the enhanced moistening over land areas peaking near 600 hPa for WS1, WS2, WS3, and WS4. This anomaly is best explained by detrainment from enhanced populations of cumulus congestus with cloud tops near the freezing level (e.g., Johnson et al. 1999). It is thus postulated that the ISCCP regimes have a more frequent occurrence of midlevel and congestus clouds over land domains for several of the weather states, consistent with the general findings of Casey et al. (2007), though further investigation is again necessary to prove this hypothesis.

4. Applications

The following section briefly illustrates two potential uses of the total diabatic heating composites derived for the ISCCP regimes. A simple linear combination of the ensemble average heating profile from all domains, weighted by the relative frequency of occurrence at each grid point for each weather state, is used to reconstruct a full four-dimensional total heating field. Each of the projects below is the subject of additional work by the various coauthors and will be explored in more detail in forthcoming publications.

a. Comparisons to CAM4 heating

The vertical structure of the annual-average, zonal-mean diabatic heating from 1984 to 2007 is shown in Fig. 9 for the ISCCP reconstruction and CAM4 total heating (i.e., the sum of moist processes, vertical diffusion,
longwave, and shortwave heating rates) for the simulation described in section 2. Corresponding difference fields (ISCCP – CAM4) are shown in the rightmost panel of Fig. 9, where warm colors indicate greater heating (or less cooling) for ISCCP. Overall, CAM4 agrees with the magnitude and extent of the ISCCP heating from deep tropical convection (especially within the ITCZ at 7.5°N), though ISCCP produces more heating (or not enough cooling) throughout the midtroposphere in much of the subtropics.

Representative maps of the annual average, ISCCP-derived and CAM4 heating (not shown) indicate that the ISCCP reconstruction matches the large-scale patterns of heating and cooling predicted by CAM4 at low levels. Although CAM4 produces excessive heating in regions with high elevations (e.g., East Africa and the Tibetan Plateau), large regions of the tropics and subtropics have small differences (<0.25 K day⁻¹) between the two datasets. Differences in the heating field become more pronounced in the middle levels and maximize over the marine boundary layer cloud regimes of the east Pacific and subtropical Atlantic oceans. There is also some difference along the northern and southern boundaries of the domain (i.e., 35°N and 35°S) in the Indian and Pacific Oceans. These regions are likely influenced by midlatitude systems (including nimbostratus clouds associated with warm and cold fronts) that are not accurately depicted in the extended low-latitude ISCCP weather states (e.g., Oreopoulos and Rossow 2011).

The largest inconsistencies between CAM4 and ISCCP generally occur over the regions with the greatest frequency of occurrence for the fair-weather cumulus regime. The notable difference in the average heating profiles for WS8 over land and ocean (Fig. 6e) suggest it may be appropriate to apply separate profiles during the linear combination over the respective areas in lieu of using the all domain composite, as is done for the other regimes. The WS8 ocean profile has cooling at middle levels, characteristic of clear sky, and would help offset the differences between CAM4 and ISCCP. Additional reconstructions using separate lookup profiles for WS8 over land and ocean were performed and the peak geographic difference was reduced by approximately 1 K day⁻¹ or 50% of the original value (not shown). The improved performance over oceans comes at the expense of larger differences over land, however, as the corresponding WS8 profile has moderately strong heating of 2–3 K day⁻¹ at middle levels. The difference between the realizations using different lookup profiles presents a range of uncertainty for our estimates, and the large-scale sensitivities and dynamical response of GCMs forced with these heating variations is the subject of future work.

b. Heating anomalies during the MJO

Despite continued advances in modeling, GCMs continue to struggle with reproducing the salient features of the MJO (Zhang et al. 2006; Li et al. 2009; among others). Changes to the shape and magnitude of the diabatic heating profiles produced by convective parameterizations have a significant effect on a GCM’s ability to simulate the MJO. Recent studies have underscored the
importance of both horizontal and vertical heating variations, with low-level heating from shallow convection thought to induce large-scale moisture convergence and preconditioning the environment for MJO initiation (e.g., Mu and Zhang 2008; Li et al. 2009; Jia et al. 2010). Lappen and Schumacher (2012) produced better simulations of the MJO in CAM4 when forcing the model with a realistic horizontal and vertical distribution of latent heating derived from the TRMM PR. They speculated their simulations would see continued improvement with the addition of low-level heating anomalies associated with shallow convection and nonprecipitating clouds that the TRMM PR cannot detect.

Maps of the heating anomalies for each weather state were composited by phase during strong MJO events (see section 2) using the Wheeler and Hendon (2004) index. Results for the cumulus regime (WS8) and the corresponding low-level heating anomalies sorted by MJO phase for November–April during 1983–2008 are shown in Fig. 10. Small areas of weak average positive anomalies (0.05–0.10 K day$^{-1}$) are evident at 940 hPa over central Africa during phase 4 (Fig. 10d) and grow in size and strength as they propagate eastward along the equator. The WS8 heating anomalies cover a broad area of the warm pool region during phase 1 and reach their peak intensity of 0.20–0.25 K day$^{-1}$ at that time. They eventually become indistinguishable as the signal propagates over the central Pacific Ocean during phases 3 and 4. The background heating for this regime (regardless of phase) is $\sim$1 K day$^{-1}$, meaning the associated anomalies may be as large as 20%–25% of the mean heating. A strong negative anomaly meanwhile lags the peak heating by 3–4 phases and is centered over the locations usually associated with deep convection for each stage of the MJO. A similar lag and exchange of suppressed clouds leading the convectively active regime was identified in Tromeur and Rossow (2010) using a subset of the tropical ISCCP weather states.

Vertical profiles of the anomalous total heating from all ISCCP regimes during each phase of the MJO are shown in Fig. 11 for the equatorial eastern Indian Ocean and west Pacific. Upper-level anomalies during the active MJO phases in each region (phases 2 and 3 in the Indian Ocean and phases 5 and 6 in the west Pacific) reach $\sim$1 K day$^{-1}$ at 400 hPa, which is generally consistent with the height and magnitude of MJO heating anomalies diagnosed using TRMM data (Jiang et al. 2011). The vertical structure of the diabatic heating composite is mostly upright (also consistent with TRMM retrievals) and does not show the characteristic westward tilt with height that is commonly identified in reanalysis data (Jiang et al. 2011). Although the ISCCP reconstructions

![Fig. 10. ISCCP composite heating anomalies by phase for WS8 (cumulus/shallow convection) at approximately 940 hPa for strong MJO events relative to the 6-month (November–April) 1983–2008 mean: (a)–(h) phase 1–8. Warm (cool) colors indicate greater heating (cooling) during strong MJO events compared to the climatological heating for the regime (regardless of phase).](image-url)
are capable of diagnosing the heating contributions from predominantly shallow boundary layer clouds (WS8), these anomalies are located far in advance of the deep convective core and appear unattached to the main upper-level heating signature.

Considering the idealized anomalous tilted heating structure of the MJO, one might expect midlevel convection and cumulus congestus (e.g., WS3) to be the principle cloud regime found throughout the east-to-west transition from shallow to deep convection. However, anomalous heating contributions from WS3 do not occur in the transition phases and are instead located along the northern and southern peripheries of the same longitudes as the mid- and upper-level heating anomalies from deep convection comprising WS1 and WS2 (not shown). This result is consistent with Tromeur and Rossow (2010), who showed the equivalent WS3 frequency of occurrence was mostly insensitive to MJO phase and Riley et al. (2011), who demonstrated that the cumulus congestus mode was relatively weak and aligned with (rather than preceding) the location of wide, deep precipitating systems using CloudSat data. Furthermore, the quick transition from WS8 to WS1 during initiation is consistent with the rapid onset of moistening and upward motion during early development of the MJO (e.g., Tromeur and Rossow 2010). Further evaluation of the MJO response in CAM4 from including the effects of the nonprecipitating regime (WS8) or total heating composites from all the weather states will be the subject of future work.

5. Summary and discussion

This study has created representative profiles of the apparent heat source $Q_1$ and moisture sink $Q_2$ for a number of unique cloud populations from a 25-yr subset of the ISCCP dataset. Profiles were created by compositing calculations of $Q_1$ and $Q_2$ derived from field campaign sounding observations across a wide variety of tropical and subtropical domains according to commonly occurring cloud mixtures (or “weather states”). While the ISCCP regimes describe populations of clouds occurring within a larger domain ($2.5^\circ \times 2.5^\circ$) and inevitably contain overlap, the composite profiles for each weather state were unique and highlight the importance of considering all components (latent, radiative, and eddy sensible heating) when determining characteristics of the total diabatic heating.

The heating profiles were well explained by the convective properties and types of clouds within each regime. The weather state characteristic of intense MCSs (WS1) had the strongest heating with a top-heavy profile owing to the large stratiform rain contribution and anvil area observed in previous work. The heating profiles showed a gradual transition to weaker values and lower heights as the convective intensity decreased (e.g., WS2
and WS3). Despite being primarily associated with thin cirrus, WS4 had the largest heating at low and middle levels. It was suggested that the low-level feature is a consequence of the redistribution of heat by cumulus clouds commonly observed with cirrus in the tropics (e.g., Jakob et al. 2005; Schumacher et al. 2008), while the midlevel feature may be from anvil clouds that otherwise are classified as WS4 based on their histograms of cloud-top pressure and optical thickness. WS5, WS6, and WS7 are all responsible for significant cooling and mainly occur over the eastern ocean basins. The regime describing mostly nonprecipitating, shallow cumulus (WS8) had weak heating near the surface with different estimates of heating or cooling aloft depending on whether the regime was present over land or ocean. Other regimes had similar total heating characteristics for land and ocean domains. The ensemble average $Q_1$ profiles for each regime were generally consistent with previous work identifying heating for specific cloud types (e.g., Nitta and Esbensen 1974; Schumacher et al. 2008).

Profiles of the apparent moisture sink showed intense drying for WS1, with weaker drying for the remaining convectively active weather states. The marine boundary layer regimes showed moistening of the lower atmosphere due to vertical eddy transport of near-surface water vapor and detrainment aloft without much loss from precipitation. Additional midlevel moistening was most pronounced for WS2 and WS3 over land, suggesting enhanced detrainment and more frequent midlevel and congestus clouds compared to ocean domains.

Two potential applications of the ISCCP composites were discussed. The first included a comparison and benchmark against diabatic heating produced from a long-term GCM simulation. Though neither the ISCCP nor CAM4 realization can be regarded as the truth, the general consensus among solutions provides some sense of validation for each. Differences in the strength of the heating were more significant at middle and upper levels, largely explained by the uncertainties in the heating profile for the cumulus regime (WS8). Future work plans to examine the large-scale response in a GCM to variations in these heating profiles and compare the results with other estimates of the Hadley circulation derived from reanalyses (e.g., Stachnik and Schumacher 2011).

A second application focused on the retrieval of low-level heating associated with shallow convection and the MJO. Although these cloud types are usually nonprecipitating and have a weak net latent heating effect, they produce low-level moistening and a redistribution of heating that may potentially induce large-scale moisture convergence and promote the onset of MJO initiation. Heating anomalies from WS8 were identified well in advance of the locations of deep convection, though the remainder of the regimes had anomalies over the same locations, suggesting a rapid change from shallow to more deep and organized convection. The observed ISCCP heating (in agreement with previous TRMM results) did not reproduce a vertically tilted diabatic heating structure, casting some doubt on whether the vertical tilt identified in reanalysis data is as evident in reality.

Finally, the separation of diabatic heating profiles by regime provides a unique metric for the evaluation of implicit cloud properties in climate models. Many GCMs now include an ISCCP simulator, and comparisons of the derived model properties with observations can yield new insights into the potential strengths and weaknesses of convective parameterizations and help quantify those cloud types with the largest uncertainties and potential impacts on tropical climate dynamics.

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