Observing Interannual Variations in Hadley Circulation Atmospheric Diabatic Heating and Circulation Strength

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ABSTRACT

Satellite and reanalysis data are used to observe interannual variations in atmospheric diabatic heating and circulation within the ascending and descending branches of the Hadley circulation (HC) during the past 12 yr. The column-integrated divergence of dry static energy (DSE) and kinetic energy is inferred from satellite-based observations of atmospheric radiation, precipitation latent heating, and reanalysis-based surface sensible heat flux for monthly positions of the HC branches, determined from a mass weighted zonal mean meridional streamfunction analysis. Mean surface radiative fluxes inferred from satellite and surface measurements are consistent to 1 W m\(^{-2}\) (<1%) over land and 4 W m\(^{-2}\) (2%) over ocean. In the ascending branch, where precipitation latent heating dominates over radiative cooling, discrepancies in latent heating among different precipitation datasets reach 22 W m\(^{-2}\) (17%), compared to 3–6 W m\(^{-2}\) in the descending branches. Whereas direct calculations of DSE divergence from two reanalyses show opposite trends, the implied DSE divergence from the satellite observations of atmospheric diabatic heating exhibits no trend in all three HC branches and is strongly correlated (reaching 0.90) with midtropospheric vertical velocity. The implied DSE divergence from satellite observations thus provides a useful independent measure of HC circulation strength variability. The sensitivity to circulation change is 4–5 times larger for precipitation latent heating compared to atmospheric radiative cooling in the descending branches and 20 times larger in the ascending branch. The difference in sensitivity is due to cloud radiative effects, which enhance atmospheric radiative cooling in the descending branches in response to an increase in HC strength but decrease it in the ascending branch.

1. Introduction

At the global scale, there is an energy balance between atmospheric radiative cooling and sensible and latent heat flux from the surface to the atmosphere. Underlying this global balance is a highly inhomogeneous regional distribution of diabatic heating and cooling. On average, radiative cooling dominates over heating in the extratropics and polar regions, and latent and sensible heating dominate in the tropics. To maintain a regional atmospheric energy balance, atmospheric circulations transport energy between source and sink regions. Idealized model experiments show that changes in the hemispheric distribution of atmospheric heating are accompanied by pronounced shifts in major circulation...
and precipitation patterns (Kang et al. 2008; Yoshimori and Broccoli 2009; Frierson et al. 2013). Furthermore, because of their influence on the vertical distribution of atmospheric heating/cooling, cloud radiative processes can have a profound influence on atmospheric circulation and precipitation patterns (Randall et al. 1989; Slingo and Slingo 1991; Wang and Ros sow 1998), underscoring the need for improved model representation of how clouds interact with their local environment (Stephens 2005; Bony et al. 2006; Fasullo and Trenberth 2012; Stevens and Bony 2013; Sherwood et al. 2014). Observing the linkage between changes in the regional distribution of atmospheric energy and its transport through the general circulation system is a challenge because of the time–space scales involved. Despite much progress, several questions remain, particularly with respect to how the position and strength of major circulation systems, such as the Hadley and Walker circulations, vary over time, how these variations influence the regional distribution of water and energy, and how clouds influence the distribution of atmospheric heating.

Satellite Earth radiation budget (ERB) observations have long been used to determine the total (atmosphere and ocean) poleward energy transport resulting from the equator-to-pole imbalance in net top-of-atmosphere (TOA) radiation (Vonder Haar and Suomi 1971; Trenberth and Caron 2001; Hartmann and Ceppi 2013). When combined with global atmospheric reanalysis systems, they have proved useful for quantifying the atmosphere and ocean contributions to the total energy transport (Trenberth and Solomon 1994; Trenberth and Caron 2001; Wong et al. 2014) and have led to improved understanding of energy and moisture budgets associated with the major atmospheric circulations (Trenberth and Stepaniak 2003; Back and Bretherton 2006).

Recently, studies have sought to provide evidence for expansion and/or weakening of the tropical circulation predicted to occur under global warming by global climate models (Vecchi and Soden 2007; Lu et al. 2007). These have primarily relied upon reanalysis data of meridional mass streamfunctions (Mitas and Clement 2005; Hu and Fu 2007; Johanson and Fu 2009; Stachnik and Schumacher 2011), reanalysis and radiosonde data of tropopause height (Seidel and Randel 2006, 2007; Lu et al. 2009), satellite observations of atmospheric ozone concentrations (Hudson et al. 2006), TOA outgoing longwave radiation (OLR) (Fu et al. 2006; Hu and Fu 2007), and cloud amount and precipitation rate (Zhou et al. 2011). While there is a general consensus that the tropics have expanded in recent decades, the magnitude of the expansion varies markedly across different datasets (Davis and Rosenlof 2012; Stachnik and Schumacher 2011). Discrepancies among reanalyses on how the atmospheric circulation strength has changed are even greater; even the sign of recent changes in circulation strength is uncertain (Bronnimann et al. 2009; Stachnik and Schumacher 2011).

The linkage between the distribution of atmospheric energy and circulation is described by the atmospheric heat budget. At annual time scales, the local and vertically integrated atmospheric energy budget involves a balance between the column-integrated divergence of dry static and kinetic energy and the sum of atmospheric radiative cooling, latent heat associated with precipitation, sensible heat flux. The advantage of using satellite instruments and datasets designed for climate research is that they are generally more stable with time than those used in reanalysis and thus less likely to exhibit spurious jumps and drifts. Further, satellite observations are less sensitive to model error, for example, in the representation of clouds and convection, aerosols, etc. The disadvantage is that there are few high-quality satellite records available, especially compared to the number of reanalysis datasets, and record lengths for the highest-quality radiation and precipitation datasets are shorter, thus limiting their use in trend analysis. Furthermore, satellite-based estimates of diabatic heating terms are not always independent of reanalysis, as there are advantages in combining the satellite measurements and reanalysis output for some of the required variables.

This study examines the use of satellite observations and reanalysis for determining interannual variations in the divergence of dry static and kinetic energy implied by atmospheric diabatic heating within the ascending and descending branches of the Hadley circulation (HC). We show that the implied energy divergence varies in proportion with circulation strength and thus provides a useful independent measure of large-scale tropical meridional circulation variability. We also use satellite observations to examine the influence of clouds on the atmospheric energy budget and show that robust relationships between interannual variations in cloud radiative effects and HC strength are attainable. To account for seasonal and interannual variations in the position and width of the HC, we use an averaging domain that changes monthly according to the location of the ascending and descending branches of the HC. The following section describes the satellite and reanalysis datasets used in this study. This is followed by a detailed
description in section 3 of the methodology used to determine atmospheric diabatic heating within the HC for latitudinal domains derived from a mass weighted zonal mean meridional streamfunction analysis. Detailed results of the analysis are presented in section 4 and briefly summarized in section 5 along with some concluding remarks.

2. Observations

We use atmospheric radiation and precipitation satellite data products developed for climate research and meridional wind and surface sensible heat flux data from reanalysis. The focus of this study is for the period between March 2000 and September 2012, selected because of the availability of multiple high-quality satellite precipitation datasets, and a radiation dataset that has been extensively compared with surface radiation measurements. Because of the lack of satellite-based surface sensible heat flux data over land during this period, we rely on reanalysis for sensible heat flux.

Atmospheric shortwave (SW), longwave (LW), and net radiative fluxes are determined monthly in 1° × 1° regions from the downward flux at the TOA minus the downward flux at the surface (SFC). Radiative fluxes are from the Clouds and the Earth’s Radiant Energy System (CERES) Energy Balanced and Filled (EBAF) Ed2.7 product for the TOA (Loeb et al. 2009a, 2012b) and SFC (Kato et al. 2013). CERES broadband radiation instruments aboard the Terra (launched in December 1999) and Aqua (launched in May 2002) satellites are used along with 3-hourly geostationary satellite observations that have been cross calibrated with the more accurate Moderate Resolution Imaging Spectroradiometer (MODIS) imager (Salomonson et al. 1989; Barnes et al. 1998), which is also aboard Terra and Aqua. The methodology used to enhance CERES Terra and Aqua diurnal sampling with geostationary data is described in detail in Doelling et al. (2013). Computed surface radiative fluxes are based upon radiative transfer model calculations initialized using MODIS and geostationary satellite–retrieved surface, cloud, and aerosol properties and reanalysis data for atmospheric state (for details, see Kato et al. 2013). In EBAF, CERES-derived TOA fluxes are used to constrain computed radiative fluxes so that observed TOA and computed SFC flux are internally consistent.

Three precipitation data products are considered. The Global Precipitation Climatology Project version 2.2 (GPCP v2.2) monthly precipitation data (Adler et al. 2003; Huffman et al. 2009) combine precipitation information from the Special Sensor Microwave Imager (SSM/I) aboard the Defense Meteorological Satellite Program (DMSP; United States) satellites, infrared precipitation estimates from geostationary satellites, the Atmospheric Infrared Sounder (AIRS) data from the National Aeronautics and Space Administration (NASA) Aqua, the Television Infrared Observation Satellite Program (TIROS) Operational Vertical Sounder (TOVS), and outgoing longwave radiation precipitation index (OPI) data from the National Oceanic and Atmospheric Administration (NOAA) satellites. Over land, gauge data from the Global Precipitation Climatology Centre (GPCC) of the Deutscher Wetterdienst and the Climate Prediction Center of NOAA are used. The second source of precipitation data is from the Tropical Rainfall Measurement Center (TRMM). We use the TRMM Microwave Imager (TMI) level 3 monthly 0.5° × 0.5° profiling V7 (3A12) data product, derived from the level 2 2A12 product (Kummerow et al. 2001). Also used is the TRMM and other sources monthly rainfall product (TRMM product 3B43) (Huffman et al. 2007), which combines multiple independent precipitation estimates from the TMI, Advanced Microwave Scanning Radiometer for Earth Observing System (AMSR-E), SSM/I, Special Sensor Microwave Imager/Sounder (SSMI/S), Advanced Microwave Scanning Unit (AMSU), Microwave Humidity Sounder (MHS), microwave-adjusted merged geo-infrared (IR), and monthly accumulated GPCC rain gauge data. All input microwave data are intercalibrated to the TRMM Combined Instrument (TCI) precipitation estimates (TRMM product 3B31) (Haddad et al. 1997), which is based upon TMI and the TRMM precipitation radar (PR). To convert GPCP and TRMM precipitation rates to latent heat flux a latent heat of condensation of 2.5 × 10^6 J kg⁻¹ is used.

Vertical profiles of meridional winds and surface sensible heat flux are from the Interim European Center for Medium-Range Weather Forecasts (ECMWF) ReAnalysis (ERA-Interim) (Dee et al. 2011). Meridional wind (V component of wind) is from the “monthly means of daily means” product at 37 pressure levels. Surface sensible heat flux is from the “synoptic monthly means” product. Direct calculations of the vertical integral of dry static and kinetic energy divergence and 500-hPa vertical velocity are obtained from both ERA-Interim and the Modern-Era Retrospective Analysis for Research and Applications (MERRA) version 5.2.0 (Rienecker et al. 2011) determined from four analyses per day. Total energy tendencies are obtained using calculations provided by the National Center for Atmospheric Research (NCAR 2014), derived from ERA-Interim.

To evaluate satellite-based estimates of surface radiative fluxes, surface observations over land and ocean are used. The land sites are from the Baseline Surface
Radiation Network (BSRN), the NOAA/Surface Radiation Network (SURFRAD), and the U.S. Department of Energy’s Atmospheric System Research (ASR) program. Over the ocean, surface buoy measurements maintained by the NOAA/Pacific Marine Environment Laboratories (PMEL) and Woods Hole Oceanographic Institute (WHOI) are used. The land measurements are located in relatively uniform terrain over a diverse range vegetation types and desert. The ocean buoys are primarily located near the equator but span all longitudes. A total of 37 land and 49 ocean sites are possible; however, in practice only up to 30 land and 19 ocean sites are available in any given month. There are fewer LW than SW ocean sites because all ocean buoys have a SW instrument but not all maintain a LW instrument as well.

In the comparisons with satellite estimates of surface radiation, if the monthly mean flux at a site contains less than 85% of its 1-min observations for a month, that site/month is excluded from the statistics. Over land, the uncertainty in downwelling flux at a given surface site is estimated to be approximately 5 W m\(^{-2}\) for both SW (Michalsky et al. 2011; Augustine and Dutton 2013) and LW (Philipona et al. 2001). Over the ocean, Colbo and Weller (2009) estimate a measurement uncertainty to be similar (4–6 W m\(^{-2}\)).

The monthly CERES ISCCP-D2like-Merged Ed2A product is used to assess how anomalies in the frequency of occurrence of different cloud types covary with cirrus clouds. This product provides the frequency of occurrence of different cloud types for 1° × 1° regions monthly. The cloud retrievals are based upon Minnis et al. (2011) for MODIS and Minnis et al. (1995) for the geostationary satellite instruments.

### 3. Methodology

To a good approximation, the energy budget of the atmosphere on an annual mean basis involves a balance between radiation, latent and sensible heat, and the column-integrated divergence of dry static energy and kinetic energy (Trenberth and Solomon 1994; Muller and O’Gorman 2011),

\[
R_a + LP + S = H^* ,
\]

where \(R_a\) is the net atmospheric radiative flux \((R_a = R_t - R_x)\); \(R_t\) and \(R_x\) are the net downward flux at the TOA and surface, respectively; \(P\) is the precipitation rate; \(L\) is the latent heat of condensation; \(S\) is the surface sensible heat flux; and \(H^*\) is the vertical integral of divergence of dry static and kinetic energy. Equation (1) neglects the total energy tendency term (NCAR 2014), as it contributes <1 W m\(^{-2}\) for the HC averaging domains considered in this study. Alternatively, the vertical integral of divergence of dry static and kinetic energy can be determined directly from

\[
H = \mathbf{V} \cdot \frac{1}{g} \int_0^P (s + k) \mathbf{u} \, dp ,
\]

where \(g\) is the acceleration due to gravity, \(s = c_p T + gz\) is the dry static energy, \(k\) is the kinetic energy, \(\mathbf{u}\) is the horizontal velocity, \(p\) is pressure, \(c_p\) is the specific heat capacity of air at constant pressure, \(T\) is absolute temperature, and \(z\) is height. Hereinafter, we use the symbol \(H^*\) to refer to divergence of \(s\) and \(k\) based upon the diabatic heating terms in Eq. (1) and \(H\) to refer to that from Eq. (2). We use CERES EBAF observations to determine \(R_a\), GPCP and/or TRMM precipitation datasets for \(LP\), and ERA-Interim for \(S\). Each term is evaluated over latitudinal domains corresponding to the ascending and descending branches of HC.

Equations similar to Eqs. (1) and (2) are obtained for divergence of moist static energy \((h = s + Lq, where q is specific humidity), with surface evaporation rate \(E\) replacing \(P\) in Eq. (1) (Trenberth and Solomon 1994). We restrict our analysis to \(s\) as the dry static energy transport and divergence show a greater contrast across the HC compared to \(h\) owing to a large compensation of \(s\) and \(Lq\) heat transports between the ascending and descending branches of the HC (Trenberth and Stepaniak 2003). Furthermore, direct global satellite-based observations of \(P\) are more mature compared to \(E\) (particularly over land) and therefore are less likely to exhibit spurious jumps and drifts over time.

To determine the HC branches, the mass weighted zonal mean meridional streamfunction \(\Psi\) is first determined from the zonal mean meridional velocity profile (Oort and Yienger 1996),

\[
\Psi = \frac{2\pi R \cos \theta}{g} \int_0^P \mathbf{v} \, dp ,
\]

where \(\mathbf{v}\) is the monthly zonal mean meridional velocity (defined positive north), \(p_s\) is the surface pressure, \(R\) is the radius of Earth, and \(\theta\) is the latitude. Figure 1a shows \(\Psi\) for October 2011 derived using meridional winds from ERA-Interim (Dee et al. 2011). Prominent Hadley cells appear on both sides of the equator. Latitudinal boundaries for the ascending and descending branches of the HC are determined with respect to the 650-hPa
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TABLE 1. Mean position of the three branches of the Hadley circulation for each calendar month between March 2000 and February 2012 derived from a streamfunction analysis of ERA-Interim meridional winds.

<table>
<thead>
<tr>
<th>Month</th>
<th>SH descending</th>
<th>Ascending</th>
<th>NH descending</th>
</tr>
</thead>
<tbody>
<tr>
<td>Jan</td>
<td>36.7°–22.5°S</td>
<td>22.5°–9.0°N</td>
<td>9.0°–29.9°N</td>
</tr>
<tr>
<td>Feb</td>
<td>37.8°–22.5°S</td>
<td>22.5°–9.0°N</td>
<td>9.0°–29.2°N</td>
</tr>
<tr>
<td>Mar</td>
<td>37.0°–18°S</td>
<td>18°–9.0°N</td>
<td>9.0°–29.5°N</td>
</tr>
<tr>
<td>Apr</td>
<td>35.2°–13.5°S</td>
<td>13.5°–10.5°N</td>
<td>10.5°–30.1°N</td>
</tr>
<tr>
<td>May</td>
<td>30.4°–7.5°S</td>
<td>7.5°–16.5°N</td>
<td>16.5°–34.1°N</td>
</tr>
<tr>
<td>Jun</td>
<td>29.2°–6.0°S</td>
<td>6.0°–21.0°N</td>
<td>21.0°–36.6°N</td>
</tr>
<tr>
<td>Jul</td>
<td>28.6°–3.0°S</td>
<td>3.0°–24°N</td>
<td>24°–42.7°N</td>
</tr>
<tr>
<td>Aug</td>
<td>29.0°–3.0°S</td>
<td>3.0°–25.5°N</td>
<td>25.5°–44.1°N</td>
</tr>
<tr>
<td>Sep</td>
<td>30.3°–4.5°S</td>
<td>4.5°–22.5°N</td>
<td>22.5°–43.2°N</td>
</tr>
<tr>
<td>Oct</td>
<td>32.1°–7.5°S</td>
<td>7.5°–18.0°N</td>
<td>18.0°–38.4°N</td>
</tr>
<tr>
<td>Nov</td>
<td>33.1°–12.0°S</td>
<td>12.0°–13.5°N</td>
<td>13.5°–35.2°N</td>
</tr>
<tr>
<td>Dec</td>
<td>34.7°–19.5°S</td>
<td>19.5°–10.5°N</td>
<td>10.5°–31.4°N</td>
</tr>
</tbody>
</table>

Fig. 1. (a) Mass weighted zonal mean meridional streamfunction for October 2011 computed using ERA-Interim meridional velocities. Solid lines show boundaries separating Hadley cell ascending and descending branches. (b) Frequency of pressure level at which minimum (SH) and maximum (NH) monthly streamfunction values are observed over the tropics (30°S–30°N) for all months between March 2000 and February 2010.

pressure level, the level most representative of the actual maximum and minimum tropical $\Psi$ for all of the months considered (Fig. 1b). Latitudes corresponding to the maximum $\Psi$ in the northern tropics (0°–30°N) ($\Psi_{\text{max}}$) and the minimum negative $\Psi$ in the southern tropics (0°–30°S) ($\Psi_{\text{min}}$) define the Northern Hemisphere (NH) and Southern Hemisphere (SH) Hadley cell center locations, respectively. Latitudes equatorward of the Hadley cell centers correspond to the ascending branch of the HC. Descending branches are defined from the Hadley cell centers to the first poleward latitudes along the 650-hPa pressure level at which $\Psi$ changes sign from positive to negative in the NH and from negative to positive in the SH. Table 1 provides the mean position of the three branches of the HC for each calendar month between March 2000 and February 2012.

4. Results

In this section, we examine the use of satellite and reanalysis data for quantifying the mean and interannual variability of the atmospheric heat budget for the ascending and descending branches of the HC. We first assess the use of meridional streamfunctions for defining averaging domains that track the monthly locations of the ascending and descending branches of the HC. We then show the mean and variability in each of the terms in Eq. (1) for each HC branch, compare different satellite-based precipitation products, and validate CERES surface radiative fluxes with ground observations. We show that interannual variations in atmospheric energy divergence inferred from satellite observations vary in proportion to interannual variations in HC strength and thus provide a useful independent measure of large-scale tropical meridional circulation variability. Finally, we evaluate how clouds influence covariability between interannual variations in HC strength and radiation at the TOA, within the atmosphere, and at the surface.

a. Hadley circulation latitudinal boundaries

The HC boundaries track the position of tropical convection and subtropical subsidence, as illustrated in Figs. 2a–d, which show the regional distribution of SW and LW cloud radiative effect (CRE), defined as the clear-sky minus all-sky TOA outgoing radiative flux difference. The ascending branch contains areas of maximum LW CRE associated with major convective regions over land (Africa, southern Asia, and South America) and ocean (intertropical convergence zone, Indian Ocean, and western tropical Pacific Ocean). The ascending branch also includes areas of subsidence in the eastern Pacific Ocean, corresponding to the downward branch of the Walker circulation. The descending branches bound areas of weak LW and SW CRE from more pronounced cloud radiative effects in the mid-latitudes. In the winter hemisphere the Hadley cell is stronger and located closer to the equator and the descending branch is wider compared to other seasons.
The close relationship between LW CRE and the HC is examined further in Figs. 3a,b. The monthly mean zonal average LW CRE exhibits robust maxima and minima associated with the ascending and descending branches of the HC, as indicated by the open circles in Fig. 3a. Prominent LW CRE maxima are also associated with the NH and SH midlatitude storm tracks. The latitude positions of the LW CRE extrema are determined as follows: In each hemisphere the midlatitude maximum position is first located from the maximum LW CRE poleward of 30°. The subtropical minimum position is determined from the minimum LW CRE between the equator and the midlatitude maximum, and the deep tropical maximum position is located at the maximum LW CRE equatorward of the subtropical minima. Figure 3b examines the relationship between the locations of tropical zonal mean LW CRE extrema and the HC latitude boundaries derived from the mass streamfunction analysis described in section 3 for each calendar month using data from March 2000 to February 2010. Minima in LW CRE fall approximately in the middle of the descending branches of the HC and correspond to the latitude where subsidence reaches a maximum, while the maximum LW CRE lies within the ascending branch and corresponds to the latitude of maximum convection. The variability in the descending branches is smallest in the wintertime when the HC is strongest, as is evident from the standard deviation in climatological mean positions represented by the error bars in Fig. 3b. In the ascending branch, the maximum LW CRE lies to the north of the equator between April and November and to the south in February and March. During transition months (March and December), the maximum LW CRE can occur on either side of the equator, explaining why the variability is larger during those months.

Interannual variations in the latitude positions of LW CRE extrema also exhibit a dependence on ENSO. Figure 4a shows the 12-month running average of the latitudinal width between NH and SH subtropical minima in LW CRE. Also provided for reference is a time series of the multivariate ENSO index (MEI) (Wolter and Timlin 1998) in Fig. 4b. Comparing Figs. 4a,b, it is evident that latitudinal widths tend to be greater than average during La Niña conditions and smaller than average during El Niño conditions. This is consistent with previous studies (Seager et al. 2003; Lu et al. 2008; Stachnik and Schumacher 2011) showing tropical expansion during the cold phase of ENSO and tropical contraction during the positive phase. These studies note that changes in Hadley cell strength in response to surface warming during El Niño are associated with a strengthening and equatorward migration of the subtropical jets, resulting in a narrowing of the tropical circulation.

The interannual variation in the latitudinal width provides further support for the need to account for monthly changes in the position and width of the HC branches when determining the heat budget terms in Eq. (1). An alternate approach to stratifying climate variables according to the HC branches is to sort by the midtropospheric vertical velocity $\omega$ (Bony and Dufresne 2005; Bony et al. 2006). While this approach provides a better overall separation between ascending and
descending circulations, without further stratification it combines data in different branches of the HC into the same \( \omega \) bin, making it more difficult to relate HC strength in a given branch with the associated atmospheric diabatic heating terms. Furthermore, it may be subject to larger model error, as it requires an accurate representation of regional distribution of \( \omega \) from reanalysis as opposed to accurate zonal mean meridional velocities.

b. Interannual variations in satellite-based atmospheric energy budget terms

The mean (March 2000–February 2010) of each term in Eq. (1) averaged within each of the three branches of the HC. While latent heating in the ascending branch exceeds that in the descending branches by 70–90 W m\(^{-2}\), atmospheric radiative cooling and sensible heat flux are both larger in the descending branches, but only by 15–25 W m\(^{-2}\) and 5–10 W m\(^{-2}\), respectively. In contrast, \( H^* \) is remarkably similar for the SH and NH descending branches (within 0.3 W m\(^{-2}\) on average), so that convergence of dry static energy (DSE) is symmetric on either side of the HC. Interannual variations in \( LP \) are much larger than those in \( Ra \) and \( S \), particularly in the ascending branch. As a result, fluctuations in \( H^* \) closely track those in \( LP \).

While \( H^* \) exhibits interannual variations that are generally consistent with expectation, absolute values are highly uncertain. At global scale, \( H^* \) should be zero and atmospheric radiative cooling \((-Ra)\) should balance with latent and sensible heat exchange between the surface and atmosphere. However, recent studies have shown that there is a substantial 10–15 W m\(^{-2}\) imbalance...
in observation-based estimates of the global atmospheric and surface energy budgets, prompting either upward adjustments to latent heating (and hence global precipitation) and surface sensible heat flux (Kato et al. 2011; Stephens et al. 2012; Wild 2012) or downward adjustments to atmospheric radiative cooling (Trenberth et al. 2009). The stated uncertainty in P from Adler et al. (2012) is ±7%, corresponding to a ±6 W m⁻² uncertainty in global mean LP. Based upon the propagation of error analyses in Loeb et al. (2009a) and Kato et al. (2013), we estimate a 1σ uncertainty in Rs of ±10 W m⁻². From Stephens et al. (2012), the uncertainty in S is estimated to be ±7 W m⁻², mainly due to substantial uncertainty over land.

Figures 7a,b shows interannual variations in global mean Rs + S, LP, and H*. The average atmospheric energy imbalance (Fig. 7b) is −13.7 W m⁻², with a standard deviation of 0.64 W m⁻². Departures from the mean imbalance reach 1.6 W m⁻² between November 2004 and October 2005, and −1.3 W m⁻² between July 2011 and June 2012. These 2σ departures are associated with extrema in LP (Fig. 7a) and occur in relatively weak ENSO conditions. Interestingly, during stronger ENSO conditions, Rs + S and LP track one another closely and H* remains close to its overall average. The energetics approach thus appears to capture global interannual variations best during moderate to strong ENSO events.

c. Evaluation of satellite observations

We compare LP from GPCP v2.2, TRMM_3_A12, and TRMM_3B43 precipitation rates for the three branches of the HC and 30°S–30°N in Table 2. The three data products agree to 3 W m⁻² (5%) in the SH descending branch, 6 W m⁻² (12%) in the NH descending branch, and 22 W m⁻² (17%) in the ascending branch. For 30°S–30°N, the spread is 12 W m⁻² (14%), with GPCP between the two TRMM datasets, exceeding TRMM_3A12 by 7 W m⁻² and falling below TRMM_3B43 by 5 W m⁻².

In a more extensive comparison involving several precipitation datasets, Adler et al. (2012) found similar results, with relative differences in precipitation rates of 10%–15% over tropical ocean areas of significant rainfall. Temporal variations in the difference between GPCP and TRMM_3A12 LP values are shown in Figs. 8a–d. The largest departures from the average difference (dashed line) occur in the ascending branch (Fig. 8b), reaching −3 W m⁻² in 2010 and 2 W m⁻² in 2007. In the SH descending branch (Fig. 8a) there is a −2 W m⁻² departure from the average difference in 2007, whereas differences remain within 1.5 W m⁻² of the mean difference in the NH descending branch.

As Rs is determined from SW and LW radiation at the TOA and surface, uncertainties in Rs variations are determined separately for TOA and surface. Loeb et al. (2007, 2012a) performed an extensive analysis of
interannual variations in TOA reflected SW and emitted LW through comparisons with other instruments [e.g., Sea-viewing Wide Field-of-view Sensor (SeaWiFS), MODIS, and AIRS] and found consistency in monthly anomalies at the 0.2 W m$^{-2}$ level (1σ). Figures 9a–d and Table 3 compare CERES estimates of surface radiation with surface observations over land and ocean. Deseasonalized monthly mean anomalies are computed from the difference between the monthly mean in a given year and the mean for the same month for all available years. The EBAF surface monthly mean fluxes are determined from $1^\circ \times 1^\circ$ latitude–longitude grid boxes that are coincident with all available surface sites with complete temporal sampling over the month. CERES EBAF surface fluxes fall within the uncertainty of the ground site measurements, and exhibit excellent consistency in monthly variability (Fig. 9). The correlation coefficient in monthly anomalies is 0.95 or greater except for downward SW radiation over ocean, where $r$ is 0.9. Importantly, there are no drifts in the EBAF-SFC surface fluxes relative to the ground measurements and the satellite and surface time series show consistent fluctuations across all ENSO states during the observation period.

d. Comparison of mean and variability in $H$ and $H^*$ from satellite observations and reanalysis

Table 4 compares satellite observed and reanalysis-based averages of the terms in Eqs. (1) and (2) for the HC branches defined in Table 1, the tropics, and the globe for March 2000–February 2010. The spread in $R_\alpha$ in the ascending branch reaches 7 W m$^{-2}$ but otherwise remains <5 W m$^{-2}$. Similarly, $S$ differs by <4 W m$^{-2}$ between ERA-Interim and MERRA. In contrast, $LP$ differs by as much as 27 W m$^{-2}$ in the ascending branch, with GPCP lower than reanalysis. As a result, satellite-based $H^*$ is at least 20 W m$^{-2}$ smaller than either $H$ or $H^*$ from reanalysis in the ascending branch. The $H^*$ values in the descending branches are within 5 W m$^{-2}$ of one another but are up to 12 W m$^{-2}$ lower than $H$ from ERA-Interim and MERRA in the SH descending branch and up to 33 W m$^{-2}$ lower in the NH descending branch. The reason for the pronounced hemispheric asymmetry in $H$ in the descending branches is unclear. Nevertheless, globally, the $H$ values come closest to energy balance.

In spite of the large global imbalance in satellite observed $H^*$ (Fig. 7b and Table 4), interannual variations

![Diagram showing the 12-month running average of the heat budget terms averaged over the ascending and descending branches of the HC.](image-url)
in $H^*$ are robust. Figures 10a–c compare interannual anomalies in $H^*$ inferred from anomalies in $R_a$ from CERES, $LP$ from GPCP or TRMM_3A12, and $S$ from ERA-Interim, with anomalies in $H$ from ERA-Interim and MERRA reanalyses. Anomalies in $H$ from direct calculations exhibit marked trends in all three branches. For ERA-Interim, there is a marked implied increase in HC circulation strength with time as $H$ increases in the ascending branch (increased divergence) and decreases in the descending branches (increased convergence). For MERRA there is no trend in the ascending branch, a weak negative trend in the SH descending branch, and a strong positive trend in the NH descending branch. In contrast, anomalies in $H^*$ based on satellite observations show no trend in all three branches.

To examine consistency in year-to-year variability, Table 5 compares correlation coefficients among the different datasets before and after removing any linear trends in the data. Correlations between the satellite and reanalysis results improve substantially after linear trends in the data are removed. They are highest in the ascending branch, reaching 0.82 when compared with MERRA, and are low in the SH descending branch. This is also the case when the two reanalyses are compared against one another. Consistency between the satellite results and MERRA reach 0.69 in the NH descending branch but are <0.5 when compared to ERA-Interim. In contrast, correlation coefficients between the two satellite-based results are >0.9 in all three branches. We note that, while different satellite precipitation datasets are used, $R_a$ and $S$ are the same in both cases, which increases the correlation.

e. Relationship between $H$ and midtropospheric vertical velocity

Changes in energy transport divergence by the mean circulation such as the Hadley cell can be decomposed into horizontal and vertical advection components (Muller and O’Gorman 2011; Brient and Bony 2012). In the zonal mean, the vertical advective component dominates (Muller and O’Gorman 2011), so that changes in the divergence of energy transport depend upon changes in mean vertical velocity (dynamical component), mean dry static stability (thermodynamic component), and their covariance. Muller and O’Gorman (2011) showed that a reasonable approximation is to assume changes in the dynamical component are proportional to the midtropospheric vertical velocity (e.g., at 500 hPa).

Figures 11a–c show time series of anomalies in $H^*$ and $\omega$ at 500 hPa, where $\omega$ is vertical velocity. Here, $\omega$ from both ERA-Interim and MERRA are considered. As $\omega$ is defined as positive downward, positive (negative) anomalies in $\omega$ correspond to increased subsidence (ascent). The results show that anomalies in $H^*$ and $\omega$ exhibit a close relationship with ENSO, particularly in the ascending branch. During the El Niño event in 2010, there is a pronounced maximum in $H^*$ (increased divergence) and minimum in $\omega$ (increased ascent). At the same time, there is a local minimum in $H^*$ (increased convergence) and a local maximum in $\omega$ (increased subsidence) in the NH descending branch. Local maxima in $H^*$ and local minima in $\omega$ in the ascending branch are also apparent during the weaker El Niño events in 2002–03, 2005, and 2007. During the 2008/09 La Niña event, a notable decrease in $H^*$ (reduced divergence) and a decline in upward vertical velocity occur in the ascending branch. Oort and Yienger (1996) observed a similar relationship between vertical velocity and ENSO, and they further showed that the Walker circulation weakens during El Niño and strengthens during

<table>
<thead>
<tr>
<th>TABLE 2. Average and standard deviation in $LP$ for SH descending, ascending, and NH descending branches of HC and 30°S–30°N based upon precipitation rates from GPCP v2.2, TRMM_3A12, and TRMM_3B43. Values in parentheses correspond to std dev of annual means. The period of coverage is March 2000–September 2012.</th>
</tr>
</thead>
<tbody>
<tr>
<td>SH</td>
</tr>
<tr>
<td>---</td>
</tr>
<tr>
<td>GPCP v2.2</td>
</tr>
<tr>
<td>TRMM_3A12</td>
</tr>
<tr>
<td>TRMM_3B43</td>
</tr>
</tbody>
</table>

Fig. 7. The 12-month running average of global mean (a) $R_a + S$ (black) and $LP$ (red) and (b) $H^*$. 

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La Niña. A decrease in $H^*$ in the ascending branch also occurs during the 2010/11 La Niña event, one of the strongest on record, but the magnitude of the decrease is roughly half that in 2008/09. The negative anomalies in $H^*$ during these La Niña events are associated with reduced precipitation over the Indian and Pacific Oceans compared to the 10-yr baseline climatology (not shown). The reduction in $H^*$ is less pronounced during the 2010/11 event because convective precipitation over the warm pool and South America is greater compared to 2008/09.

Correlation coefficients between anomalies in $H^*$ and $\omega$ before and after any linear trends are removed are provided in Table 6. In general, correlation coefficients improve after linear trends are removed. For $H^*$ from the satellite observations, correlations with $\omega$ reach $-0.90$ in the ascending branch, $-0.62$ in the SH descending branch, and $-0.77$ in the NH descending branch. In all but one case, correlation coefficients are significant at the 95% significance level. Thus, the dynamical component accounts for much of the variability in $H^*$. For the reanalyses, correlations in the descending branches show mixed results. For MERRA, there is no correlation between $H$ and $\omega$ in the SH descending branch both before and after linear trends are removed. For ERA-Interim, the correlation coefficient changes from $-0.75$ before removing linear trends to $-0.12$ afterward. Both reanalyses show strong correlation coefficients between $H$ and $\omega$ in the ascending branch.

Figures 12a–c show the slope of least squares linear regression fits between deseasonalized monthly mean anomalies in each of the terms in Eq. (1) and $\omega^*$, where $\omega^*$ is vertical velocity determined from $\Psi$ (Oort and Yienger 1996) at the 650-hPa pressure level (section 3),

$$\omega^* = \frac{g}{2\pi R^2 \cos\theta} \left| \frac{\Delta \Psi}{\Delta \theta} \right|. \quad (4)$$

In this definition, $\omega^*$ is defined positive downward in the descending branches and positive upward in the ascending branch. In the descending branches, an increase in $\omega^*$ is associated with enhanced atmospheric radiative cooling (decrease in $R_a$), a reduction in latent heating $LP$, and a negligible change in $S$. In the ascending branch, an increase in $\omega^*$ is associated with an increase in $LP$ but negligible changes in $R_a$ and $S$. Accordingly, divergence (convergence) of dry static energy increases with $\omega^*$ in the ascending (descending) branch. Interestingly, even though the mean $LP$ is a factor of 2–3 larger in the ascending branch compared to the descending branches (Figs. 6d,e) because of the large-scale convergence of moisture into the ascending branch, the covariability of $LP$ and circulation strength $\omega^*$ is approximately the same in all three branches (Fig. 12b). Furthermore, the slope of the least squares regression fit between $LP$ and $\omega^*$ is 4–5 times greater than that between $R_a$ and $\omega^*$ in the descending branches and roughly
20 times greater in the ascending branch. The larger sensitivity in LP to $\omega^*$ in the ascending branch is due to a reduction in $R_a$, which is the result of a cancellation between cloud radiative effect and clear-sky LW change with $\omega^*$ (see section 4f for details).

The smaller contribution to diabatic heating by atmospheric radiation compared to latent heating has been discussed in several modeling studies (Slingo and Slingo 1988; Randall et al. 1989; Fowler and Randall 1994; Sherwood et al. 1994; Tian and Ramanathan 2003; Bergman and Hendon 2000). These studies show that, while radiative contributions (e.g., due to clouds) appear to have a smaller direct influence on diabatic heating compared to latent heating, the two processes are intricately linked: radiatively induced diabatic heating alters the local horizontal distribution of atmospheric heating, which both directly influences the local circulation and indirectly alters convective latent heating, causing further changes in diabatic heating and circulation (Sherwood et al. 1994; Wang and Rossow 1998). In general circulation model experiments in which cloud radiative effects are turned off, the large-scale circulation (Hadley and Walker circulations) is much weaker and the atmospheric temperature, moisture, and precipitation structure is very different (e.g., Randall et al. 1989; Slingo and Slingo 1991; Sherwood et al. 1994).

A seemingly paradoxical result is that, while changes in global LP oppose those in global $R_a$ (Fig. 7a) (Allen and Ingram 2002; Stephens and Ellis 2008; Allan 2009), the two vary in phase with one another locally with variations in HC strength (Figs. 12a,b). When global monthly anomalies in $R_a$ and LP are compared with those in the three HC branches, approximately 20% of the global LP variability is explained by LP variations in the ascending branch and 20%–25% of the global variability in $R_a$ is explained by $R_a$ variations in the descending branches. A stronger HC thus contributes to enhanced global radiative cooling and latent heating at interannual time scales. The situation differs markedly under global warming, where global radiative cooling and precipitation increase and the Hadley cell weakens and expands poleward (Held and Soden 2006; Lu et al. 2008; Stephens and Ellis 2008).

### f. Covariability of atmospheric cloud radiative effects and HC strength

To explore how radiation varies with circulation strength in the three HC branches, Fig. 13 shows the slope of linear regression fits between monthly anomalies in HW against monthly anomalies in $\omega^*$ for TOA, within-atmosphere, and surface radiation. Results for both all sky (in black) and
CRE (in red) are provided. At the TOA (Figs. 13a–c), an increase in $\omega^*$ is associated with an increase in TOA LW emission to space (cooling) in the ascending branches and a decrease (warming) in the ascending branch (Fig. 13a). These are due to variations in both clouds, as indicated by the CRE bars, and clear sky (difference between all-sky and CRE bars). Conversely, TOA absorbed SW radiation increases with $\omega^*$ in the descending branches and strongly decreases in the ascending branch (Fig. 13b), owing primarily to cloud variations. Overall, Fig. 13c shows that an increase in $\omega^*$ leads to a decrease in net TOA radiation in all three HC branches, although the decrease is not significant at the 95% confidence level. At the SFC, all-sky downward LW flux shows little sensitivity to $\omega^*$ in the descending branches and an increase with $\omega^*$ in the ascending branch (Fig. 13g). The weak sensitivity in surface LW CRE to $\omega^*$ compared to that at TOA (Fig. 13a) is because it is primarily high clouds that covary with $\omega^*$. This is illustrated in Fig. 14, which shows that, while linear regression fits between cloud frequency of occurrence and $\omega^*$ anomalies are significantly different from zero for high clouds in all three branches; the slopes are not significant for low clouds. Surface LW CRE is less sensitive to high cloud variations because of their cold temperatures. Absorbed SW radiative flux (Fig. 13h) increases with $\omega^*$ in the descending branches and strongly decreases in the ascending branch, a pattern very similar to that at TOA. In all three HC branches, SW contributions dominate at the surface, so that net surface radiative warming increases with $\omega^*$ in the descending branches and decreases in the ascending branch (Fig. 13f). Within the atmosphere, LW radiative cooling increases with $\omega^*$ in the descending branches but remains largely unchanged.

Table 3. Comparison between monthly mean CERES EBAF surface SW and LW downward radiation with surface measurements over land and ocean between March 2000 and February 2011. Here, $\sigma$ is the standard deviation of monthly mean fluxes; $\sigma$ is the standard deviation in deseasonalized monthly anomalies; $\sigma (\delta C - \delta O)$ is the standard deviation of the difference in deseasonalized monthly anomalies between CERES and the surface observations; $r$ is the correlation coefficient between CERES EBAF and surface observed deseasonalized monthly anomalies; and $(N)$ is the average number of surface sites available in a given month for the comparison.

<table>
<thead>
<tr>
<th></th>
<th>Mean (W m$^{-2}$)</th>
<th>$\sigma$(Mean) (W m$^{-2}$)</th>
<th>$\sigma$ (W m$^{-2}$)</th>
<th>$\sigma (\delta C - \delta O)$ (W m$^{-2}$)</th>
<th>$r$</th>
<th>$(N)$</th>
</tr>
</thead>
<tbody>
<tr>
<td>SW land</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Surface obs</td>
<td>180.8</td>
<td>30.1</td>
<td>6.8</td>
<td>1.8</td>
<td>0.97</td>
<td>29.2</td>
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<tr>
<td>CERES EBAF</td>
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<td>29.2</td>
<td>7.0</td>
<td>1.1</td>
<td>0.99</td>
<td>29.5</td>
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<td>25.6</td>
<td>8.0</td>
<td>1.1</td>
<td>0.99</td>
<td>29.5</td>
</tr>
<tr>
<td>CERES EBAF</td>
<td>315.8</td>
<td>27.0</td>
<td>8.0</td>
<td>3.1</td>
<td>0.9</td>
<td>17.5</td>
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<tr>
<td>SW ocean</td>
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<tr>
<td>Surface obs</td>
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<td>11.8</td>
<td>7.1</td>
<td>3.1</td>
<td>0.95</td>
<td>5.4</td>
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<td>6.1</td>
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<tr>
<td>LW ocean</td>
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<td></td>
</tr>
<tr>
<td>Surface obs</td>
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<td>8.1</td>
<td>7.8</td>
<td>2.7</td>
<td>0.95</td>
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Table 4. Average $R_c$, $L_P$, $S$, $H^*$, and $H$ for March 2000–February 2010 from observations and reanalysis for latitude positions of HC branches defined in Table 1, 30°S–30°N, and the globe. Here, “CER” = CERES; “ERAI” = ERA-Interim; $H^*$ (CGE) = column-integrated DSE and kinetic energy inferred from CERES observations of atmospheric radiation, GPCP precipitation latent heating, and ERA-Interim surface sensible heat flux.

<table>
<thead>
<tr>
<th></th>
<th>SH descending</th>
<th>Ascending</th>
<th>NH descending</th>
<th>30°S–30°N</th>
<th>Global</th>
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<tr>
<td>$R_c$(CER)</td>
<td>−115.2</td>
<td>−93.5</td>
<td>−109.5</td>
<td>−106.3</td>
<td>−108.6</td>
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<tr>
<td>$R_c$(ERAI)</td>
<td>−119.7</td>
<td>−100.7</td>
<td>−107.4</td>
<td>−110.6</td>
<td>−109.2</td>
</tr>
<tr>
<td>$R_c$(Merra)</td>
<td>−118.5</td>
<td>−96.6</td>
<td>−107.4</td>
<td>−108.9</td>
<td>−105.9</td>
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<tr>
<td>$L_P$(GPCP)</td>
<td>59.3</td>
<td>131.6</td>
<td>51.3</td>
<td>88.4</td>
<td>77.6</td>
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<tr>
<td>$L_P$(ERAI)</td>
<td>61.1</td>
<td>159.0</td>
<td>54.4</td>
<td>102.4</td>
<td>82.9</td>
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<tr>
<td>$L_P$(Merra)</td>
<td>54.8</td>
<td>156.1</td>
<td>54.6</td>
<td>101.2</td>
<td>81.0</td>
</tr>
<tr>
<td>$S$(ERAI)</td>
<td>23.1</td>
<td>16.4</td>
<td>26.4</td>
<td>20.6</td>
<td>17.5</td>
</tr>
<tr>
<td>$S$(Merra)</td>
<td>27.4</td>
<td>14.0</td>
<td>26.3</td>
<td>20.3</td>
<td>18.1</td>
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<tr>
<td>$H^*$ (CGE)</td>
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<td>54.5</td>
<td>−31.8</td>
<td>2.7</td>
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<td>$H^*$ (ERAI)</td>
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<td>74.7</td>
<td>−26.6</td>
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<td>−8.7</td>
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<tr>
<td>$H^*$ (Merra)</td>
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<td>73.6</td>
<td>−26.6</td>
<td>12.6</td>
<td>−6.8</td>
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<td>$H$(ERAI)</td>
<td>−23.9</td>
<td>77.8</td>
<td>−6.1</td>
<td>23.3</td>
<td>1.4</td>
</tr>
<tr>
<td>$H$(Merra)</td>
<td>−29.3</td>
<td>78.6</td>
<td>0.9</td>
<td>25.3</td>
<td>−0.5</td>
</tr>
</tbody>
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in the ascending branch (Fig. 13d) because of a cancel- 
lation between LW radiative cooling in clear sky and LW 
radiative heating by clouds. In the SW (Fig. 13f), ab-
sorption is weakly dependent upon $\omega^*$ in all three HC 
branches, so that within-atmosphere net radiation 
change with $\omega^*$ is very similar to that in the LW.

The results suggest that, as HC circulation strength 
increases (i.e., $\omega^*$ increases), clouds enhance atmo-
spheric radiative cooling in the descending branches 
but oppose it in the ascending branch. This enhances 
the latitudinal gradient in diabatic heating, potentially 
leading to an increase in HC strength. However, the

Table 5. Correlation coefficient between time series in Figs. 10a–c without and with linear trends in the data removed. Here, CGE = $\delta R_c$(CER) + $\delta P$(GPCP) + $\delta S$(ERA-I), CTE = $\delta R_c$(CER) + $\delta P$(TRMM_3A12) + $\delta S$(ERA-I), $\delta H$(ERA-I), and $\delta H$(MERRA) for the three branches of the HC. Bold corresponds to a significant correlation at 95% significance level.

<table>
<thead>
<tr>
<th></th>
<th>SH descending</th>
<th>Ascending</th>
<th>NH descending</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>CTE</td>
<td>ERA-I</td>
<td>MERRA</td>
</tr>
<tr>
<td>Without linear trend removed</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>CGE</td>
<td>0.93</td>
<td>0.16</td>
<td>0.14</td>
</tr>
<tr>
<td>CTE</td>
<td>0.00</td>
<td>-0.10</td>
<td>0.37</td>
</tr>
<tr>
<td>ERA-I</td>
<td>0.37</td>
<td>0.41</td>
<td>0.73</td>
</tr>
<tr>
<td>With linear trend removed</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>CGE</td>
<td>0.95</td>
<td>0.44</td>
<td>0.25</td>
</tr>
<tr>
<td>CTE</td>
<td>0.51</td>
<td>0.12</td>
<td>0.59</td>
</tr>
<tr>
<td>ERA-I</td>
<td>-0.13</td>
<td>0.78</td>
<td>0.27</td>
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</table>
cloud response also influences convective latent heating in the tropics by altering the dry and moist static stability of the atmosphere (Wang and Rossow 1998). Because of the close coupling between CREs, moist convection, and circulation strength and lacking information on the atmospheric circulation response to the cloud radiative effects, it is exceedingly difficult to draw conclusions about the magnitude and even the sign of the feedback between clouds and HC strength from satellite observations alone. These are best determined through combined use of model simulations in which CREs can be turned on and off (e.g., Slingo and Slingo 1988; Randall

![Fig. 11. The 12-month running average of anomalies in $H$ and $\omega$ for the (a) NH descending, (b) ascending, and (c) SH descending branches of the HC. Note that the y axis scale for $\omega$ anomalies is reversed.](image)

| Table 6. Correlation coefficient between time series in Figs. 11a–c without and with linear trends in the data removed. Here, CGE = $\delta R_e$(CER) + $\delta P$(GPCP) + $\delta S$(ERA-I), CTE = $\delta R_e$(CER) + $\delta P$(TRMM_3A12) + $\delta S$(ERA-I), $\delta H$(ERA-I), and $\delta H$(MERRA) for the three branches of the HC. Bold corresponds to a significant correlation at 95% significance level. |
|---|---|---|---|---|---|---|---|---|
| SH descending | Ascending | NH descending |
| | $\omega$(ERA-I) | $\omega$(MERRA) | $\omega$(ERA-I) | $\omega$(MERRA) | $\omega$(ERA-I) | $\omega$(MERRA) |
| $H$(CGE) | -0.39 | -0.61 | -0.44 | -0.90 | -0.53 | -0.28 |
| $H$(CTE) | -0.31 | -0.54 | -0.59 | -0.80 | -0.35 | -0.44 |
| $H$(ERA-I) | -0.72 | -0.16 | -0.88 | -0.11 | -0.75 | 0.71 |
| $H$(MERRA) | 0.012 | -0.18 | -0.54 | -0.73 | 0.56 | -0.89 |

Without linear trend removed

| $H$(CGE) | -0.50 | -0.55 | -0.65 | -0.90 | -0.77 | -0.66 |
| $H$(CTE) | -0.55 | -0.62 | -0.73 | -0.81 | -0.74 | -0.64 |
| $H$(ERA-I) | -0.68 | -0.12 | -0.90 | -0.46 | -0.12 | -0.13 |
| $H$(MERRA) | 0.33 | -0.13 | -0.68 | -0.74 | -0.57 | -0.58 |

With linear trend removed
et al. 1989; Sherwood et al. 1994; Wang and Rossow 1998) and observations of the covariability between clouds, circulation, and the atmospheric heat budget, which can be used to test the realism of the mode simulations.

5. Summary and conclusions

Satellite observations and reanalysis data are used to evaluate observation-based interannual variations in the atmospheric heat budget terms and their relationship with the circulation within the ascending and descending branches of the Hadley circulation (HC). The column-integrated divergence of dry static energy (DSE) and kinetic energy is inferred from the sum of satellite-based observations of atmospheric radiation, precipitation latent heating, and reanalysis-based surface sensible heat flux. These quantities are evaluated over latitudinal domains corresponding to the monthly positions of the HC branches, determined from a mass weighted zonal mean meridional streamfunction analysis based on ERA-Interim vertical profiles of zonal mean meridional velocity.

The consistency between ERA-Interim HC branch latitudinal boundaries and CERES satellite observations of SW and LW cloud radiative effect (CRE) is evaluated. Overall, the HC boundaries follow the seasonal migration of tropical convection and subtropical subsidence observed in CERES data. The latitude of maximum subtropical subsidence in each hemisphere, determined from the position of minimum zonal mean LW CRE, lies approximately in the middle of the descending branches of the HC. Similarly, the tropical maximum zonal mean LW CRE, corresponding to the latitude of maximum convection, lies well within the ascending branch. The latitudinal distance between southern and Northern Hemisphere LW CRE subtropical minima is also found to correlate with the multivariate ENSO index (MEI), with greater latitudinal widths occurring during La Niña conditions and smaller widths during El Niño conditions.

At most longitudes within the HC branches, the dominant terms in the atmospheric energy budget are latent heating due to precipitation, radiative cooling, and divergence of DSE, with sensible heat playing a smaller role. In the ascending branch, latent heating dominates over radiative cooling, while the opposite is true in the descending branches. As a result, divergence (convergence) of DSE occurs in the ascending (descending) branches of the HC. Year-to-year fluctuations in latent heating exceed those in radiative cooling and sensible heating in all three HC branches and therefore accounts for most of the interannual variability in the divergence/convergence of DSE.

At global scales, the atmospheric heat balance should be between atmospheric radiative cooling and latent and sensible heating. While interannual variations in these terms track one another reasonably well, the observed atmospheric energy budget has a significant imbalance. Consistent with earlier studies, we find an imbalance in the global mean atmospheric energy budget of $\approx 14$ W m$^{-2}$, suggesting that either global radiative cooling is overestimated and/or latent and/or sensible heating are underestimated. To probe deeper, comparisons between different precipitation datasets and between satellite and surface radiation observations are performed. Precipitation rates from GPCP v2.2, TRMM_3A12, and TRMM_3B43 datasets are found to agree to 3 W m$^{-2}$ (5%) in the SH descending branch, 6 W m$^{-2}$ (12%) in the NH descending branch, and 22 W m$^{-2}$ (17%) in the ascending branch. For 30°S–30°N, the spread is 12 W m$^{-2}$ (14%). In contrast, temporal variations between GPCP and TRMM_3A12 $LP$ are consistent in all three branches. Mean surface radiative fluxes inferred from satellite and a network of surface measurements are consistent to 1 W m$^{-2}$ (<1%) over land and 4 W m$^{-2}$.
(2%) over ocean and show excellent consistency in interannual variability. While these results do not conclusively explain why there is an observed atmospheric energy imbalance, they do point to a need for improved precipitation rates in areas of heavy precipitation (e.g., ascending branch of HC).

When 10-yr averages in satellite-based diabatic heating terms are compared with those from reanalysis, large discrepancies in precipitation latent heating are found in the ascending branch, with reanalysis exceeding GPCP by as much as 27 W m$^{-2}$. Differences in LP in the descending branches and in net atmospheric radiative flux in all three branches are smaller than 7 W m$^{-2}$. Further, the DSE divergence computed directly from ERA-Interim and MERRA differ by as much as 33 W m$^{-2}$ compared to values inferred from the atmospheric diabatic heating terms in the NH descending branch and show a pronounced hemispheric asymmetry in the descending branches. Anomalies in DSE divergence computed from ERA-Interim and MERRA reanalyses are inconsistent in all three HC branches, with both showing appreciable trends that diverge from one another, particularly in the ascending and Northern Hemisphere (NH) descending HC branches. In contrast, satellite-based anomalies in DSE divergence exhibit no drifts or trends in all three branches. After removing linear trends in the data, correlation coefficients between the reanalyses and satellite-derived DSE divergence improve, reaching 0.82 in the ascending branch, 0.50 in the SH descending branch, and 0.70 in the NH descending branch. Anomalies in satellite DSE divergence are highly correlated with anomalies in vertical velocity at 500 hPa, reaching 0.90 in the ascending branch, and both show a close relationship with ENSO. During El Niño events, divergence (convergence) of DSE in the ascending (descending) branch increases, while the opposite occurs during La Niña events. We note that the situation may be quite different under global warming, where global precipitation increases and the Hadley cell weakens and expands poleward.

The slopes of the linear least squares regression fits between each of the terms in the atmospheric heat
budget and vertical velocity shows that atmospheric diabatic heating increases with circulation strength in the ascending branch and decreases in the descending branches. The regression slope between precipitation latent heating and vertical velocity is 4–5 times larger than that between net within-atmosphere radiation and vertical velocity in the descending branches and roughly 20 times larger in the ascending branch. The larger ratio in the ascending branch is due to cloud radiative effects, which offset a clear-sky LW radiative cooling increase with vertical velocity. As several modeling studies have noted, cloud radiative effects can also have a strong indirect influence on diabatic heating and circulation by altering the local distribution of atmospheric heating, giving rise to further increases in latent heating. The results show that, as HC circulation strength increases, clouds enhance radiative cooling in the descending branches but oppose it in the ascending branch.

Our ability to observe how the regional distribution of within-atmosphere energy and the large-scale circulation change with time is far from complete, even for the most basic features, such as how the position and strength of major circulation systems vary with time. The close relationship between divergence of DSE and vertical velocity in the three branches of the HC and the demonstrated stability of satellite-based observations of key aspects of atmospheric diabatic heating provide an attractive alternative approach to reanalysis-only methods for tracking changes in large-scale atmospheric circulation. However, this approach requires a continuation of well-calibrated satellite observations without data gaps (Loeb et al. 2009b) in the record, as well as backward compatibility between new instruments intended to extend a satellite record in order to avoid artificial jumps that could introduce spurious trends. With the upcoming Global Precipitation Measurement (GPM) core satellite set to launch in 2014 and the final copy of the CERES instrument (flight model 6) to launch in 2016, followed by a subsequent instrument [radiation budget instrument (RBI)] planned for 2021, there is some hope that we are on track toward establishing a high-quality multidecadal record of key measurements needed to observe changes in Earth’s atmospheric energy budget. Complementary measurements, such as those from the Climate Absolute Radiance and Refractivity Observatory (CLARREO; Wielicki et al. 2013), currently in preformulation phase, will further enhance the absolute accuracy and stability of atmospheric energy budget observations by enabling systematic intercalibration campaigns in orbit between CLARREO instruments and sun-synchronous and geostationary visible, infrared, and broadband instruments used to monitor changes in Earth’s atmospheric energy budget.

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