Interannual variability of monsoon precipitation and local subcloud equivalent potential temperature

Authors: John V. Hurley and William R. Boos

Affiliation:

1 - Department of Geology and Geophysics, Yale University, New Haven, CT, 06520

Correspondence to:

John V. Hurley
Department of Geology and Geophysics
Yale University
New Haven, CT 06520
email - john.hurley@yale.edu

Keywords: monsoon, interannual variability, convective quasi-equilibrium, equivalent potential temperature
The interannual variability of monsoon precipitation is described in the context of a convective quasi-equilibrium framework. Using two reanalysis products and two global precipitation datasets, we examine linear relationships between seasonal anomalies of precipitation and subcloud equivalent potential temperature ($\theta_{eb}$) local to six monsoon regions. This approach provides a single, near-surface, thermodynamically relevant variable over both land and ocean, extending previous studies of interannual monsoon variability that emphasized ocean surface temperatures. After removing the variability linearly associated with an index of the El Nino-Southern Oscillation, positive monsoon precipitation anomalies are shown to be associated with enhanced $\theta_{eb}$ local to and slightly poleward of the climatological $\theta_{eb}$ maximum. The variations in continental $\theta_{eb}$ local to the monsoon precipitation maxima are due mainly to variations in subcloud specific humidity, with changes in subcloud temperature having the opposite sign. Motivated by the fact that some of these subcloud humidity anomalies occur over deserts poleward of monsoon regions, the relationship of 700 hPa flow with precipitation is examined, and enhanced precipitation in several regions is found to covary with the properties of shallow meridional circulations. The implications of these results for the understanding of monsoon interannual variability are discussed.
1 INTRODUCTION

Earth’s seasonal cycle of insolation drives monsoon circulations only indirectly, with ocean and land surfaces absorbing most shortwave radiation before transferring the energy of that radiation to the overlying atmosphere via turbulent heat fluxes. It is thus no surprise that variations in the strength of monsoons have, for over a century, been associated with changes in the properties of land and ocean surfaces. For example, Blanford (1884) suggested that unusually heavy Himalayan snowfall led to drought over northwestern India, and Charney (1975) argued, in the context of the Sahel, that reductions in vegetation could increase land surface albedo and enhance atmospheric subsidence in a positive feedback. Both of these ideas relate monsoon precipitation to land-atmosphere interaction near the poleward edge of the monsoon domain.

Attention in recent years has shifted from the influence of land surface properties to the covariation of sea surface temperature (SST) with monsoon precipitation, so that SST changes are often considered to be the leading cause of monsoon interannual variability (Rasmusson and Carpenter 1983; Yang and Lau 2004). A decades-long drought in the Sahel has been linked not to Africa’s surface albedo but to changes in tropical SST (Biasutti et al. 2008; Giannini et al. 2003; Hoerling et al. 2006; Lu 2009). Australian monsoon activity is strongly correlated with Pacific SST (Taschetto et al. 2009; Taschetto et al. 2010), and interannual variations in South Asian precipitation are associated with SST changes in both the El Nino-Southern Oscillation [ENSO, (Li et al. 2007; Shukla and Paolino 1983)] and an east-west Indian Ocean dipole [IOD, (Saji et al. 1999; Webster et al. 1999)].

But even SST variations associated with ENSO, which have an especially strong relationship with rainfall in the South Asian and Australian monsoons, are linearly related to less
than half of the interannual variance of precipitation in those regions (Webster et al. 1998). This is perhaps expected since monsoons are thought to be caused by a contrast in the thermal forcing between land and ocean, and land surface thermal forcing is likely at least somewhat independent of SST. When attempting to assess the state of a monsoon’s forcing, one is thus in the position of trying to combine the SST field with some physically relevant measure of land surface properties. Land surface temperature, although well-observed, is largely unrelated to the strength of monsoon circulations, in which the phase changes of atmospheric water play a first-order thermodynamic role (Fasullo and Webster 2003). Blanford’s hypothesized relation between Himalayan snow cover and Indian precipitation has not been verified in observations (Robock et al. 2003), and the influence of soil moisture on monsoon precipitation exhibits substantial regional variation and is difficult, at best, to assess observationally (Douville et al. 2001; Godfred-Spenning and Reason 2002). While previous studies have used numerical models to examine the sensitivity of monsoon precipitation to snow cover, soil moisture, surface albedo and other parameters (Douville et al. 2001; Vernekar et al. 1995), we know of no variable representing the state of a land surface that has been shown to be strongly correlated with observed interannual variations of precipitation in multiple monsoon regions.

The goal of this paper is to examine the covariation of monsoon precipitation with a near-surface thermodynamic variable that is relevant over both land and ocean. We employ the framework of convective quasi-equilibrium (QE), in which precipitating convection is hypothesized to consume convective available potential energy (CAPE) fast enough for any changes in CAPE to be small compared to the rate of generation of CAPE by radiative cooling and surface heat fluxes (Arakawa and Schubert 1974; Emanuel et al. 1994; Neelin et al. 2008). In a strict form of QE, any changes in CAPE are assumed to be dynamically negligible, so that
temperatures within the convecting layer follow a moist adiabat tied to the equivalent potential temperature of air below the base of cumulus clouds ($\theta_{eb}$). This strict form of QE is theoretically attractive because it dramatically simplifies the vertical structure of the troposphere and ties it to a single near-surface variable, but it is clear that this is an oversimplification for many tropical atmospheric phenomena (Arakawa 2004; Brown and Bretherton 1997). Nevertheless, strict forms of QE have been used in numerous theoretical studies of monsoon dynamics (Boos and Emanuel 2008b; Chou et al. 2001; Neelin 2007; Prive and Plumb 2007a, 2007b), and have been shown to provide a decent description of the seasonal-mean state of some observed monsoons (Nie et al. 2010). In particular, Nie et al. (2010) showed that maxima of $\theta_{eb}$ in many of Earth’s regional monsoons are located in the same positions as maxima of upper-tropospheric temperature ($T_u$). This coincidence of maxima of $\theta_{eb}$ and $T_u$ is expected if moist convection couples those two variables only in a narrow monsoonal convergence zone located near the thermal maximum; moist convection and thus the coupling between $\theta_{eb}$ and $T_u$ are suppressed in the broad subsiding branch of the overturning circulation that occupies the rest of the domain.

Since $\theta_{eb}$ and $T_u$ appear to be related in this relatively simple way in seasonal climatologies of observed monsoons, it seems reasonable to ask if something can be learned by examining interannual variations of these variables. More precisely, are interannual variations in $\theta_{eb}$ linearly related to interannual variations in monsoon precipitation? And does $T_u$ covary in a similar fashion? Values of $\theta_{eb}$ will be closely tied to local SST over ocean, although they can vary independently of SST if subcloud relative humidity and air-sea temperature difference are not constant. Over land, $\theta_{eb}$ is less constrained and likely has less persistence than SST. One thus cannot regard $\theta_{eb}$ as a lower boundary condition for monsoon flow, but examination of its interannual variations may nevertheless help in understanding the processes governing monsoon
variability. In particular, since $\theta_{eb}$ is defined over both land and ocean, it allows monsoon variability to be assessed in a simple theoretical framework in which the thermodynamic state of near-surface air is represented by a single variable throughout the entire domain.

This is the first study, as far as we know, that examines covariations between precipitation in multiple monsoon regions and equivalent potential temperature on a terrain-following surface in the subcloud layer. While detailed exploration of the mechanisms responsible for these covariations is left for future work, we do have a general hypothesis that a monsoon circulation and its associated precipitation will be stronger when the meridional gradient of $\theta_{eb}$ is stronger (Emanuel 1995). Strengthening of the meridional gradient may occur through increased $\theta_{eb}$ at the location of the $\theta_{eb}$ maximum in the summer hemisphere, or by reduced $\theta_{eb}$ near the equator or in the winter hemisphere. This positive relationship between monsoon precipitation and the meridional gradient of $\theta_{eb}$ was illustrated clearly by Eltahir and Gong (1996), using observations of boundary layer moist entropy for one wet and one dry year (1958 and 1960, respectively) of the West African monsoon. The present work can thus be regarded as an extension of Eltahir and Gong (1996) to six monsoon regions using a much longer observational record.

Like Eltahir and Gong (1996), we find that monsoon precipitation does covary interannually with $\theta_{eb}$. But unlike their work, which focused on the meridional gradient of $\theta_{eb}$ between the precipitation maximum and the equatorial ocean, we show that a robust feature across all six monsoon regions is the covariation of monsoon precipitation with continental $\theta_{eb}$ near and somewhat poleward of the climatological maxima of precipitation and $\theta_{eb}$. For this reason, the scope of this manuscript is limited to examination of the association of monsoon precipitation with $\theta_{eb}$ in areas local to each monsoon region, and exploration of the relationship between precipitation and $\theta_{eb}$ in remote regions is left for future work.
We recognize that the covariation of $\theta_{eb}$ with monsoon precipitation may be more complicated than can be described by the simple QE framework used in previous theoretical studies. Our analyses do reveal such complexity, and the particular finding that monsoon precipitation covaries with $\theta_{eb}$ over the deserts located poleward of some monsoons motivates a preliminary investigation of the covariance of precipitation with the dry, shallow circulations associated with those deserts (Thorncroft and Blackburn 1999; Thorncroft et al. 2011; Trenberth et al. 2000; Zhang et al. 2008). This study also includes a brief analysis of lead-lag relationships between monsoon precipitation and $\theta_{eb}$ in one regional monsoon. The main task of this paper, though, is a straightforward documentation of the contemporaneous relationship between observed monsoon precipitation and local $\theta_{eb}$.

The next section of this paper presents the data sources and methods used for the linear regression analysis that forms the core of this work. Results of the regression analyses are presented in Section 3, which also explores both the sensitivity of results to data source and antecedent relationships in Australia. The paper ends with a short summary and discussion. An appendix presents some details concerning methodological choices.

2 DATA AND METHODS

2.1 Data sources

Global distributions of $\theta_{eb}$ were calculated using subcloud temperature ($T_b$), subcloud specific humidity ($q_b$), and subcloud pressure from both the European Centre for Medium-Range Weather Forecasts (ECMWF) Reanalysis-40 [ERA-40; (Uppala et al. 2005)] and the National Center for Environmental Prediction/National Center for Atmospheric Research (NCEP/NCAR) Reanalysis (Kalnay et al. 1996). Output for model level 57 (of 60 total levels) was used for the
ERA-40 Reanalysis and model level 27 (of 28) was used for the NCEP Reanalysis, both of which follow the terrain about 20 hPa above the surface. For both reanalyses, \( \theta_{eb} \) was calculated from monthly mean data following Emanuel (1994), where

\[
\theta_{eb} = T \left( \frac{P_{\text{po}}}{P_{\text{d}}} \right)^{\frac{R_d}{C_{pd} + C_{cl}}} \left( H \right)^{-\frac{R_v}{C_{pd} + C_{cl}}} \exp \left( \frac{L_v r}{(C_{pd} + C_{cl}) T} \right). \tag{1}
\]

\( T \) is temperature, \( P_{\text{po}} \) and \( P_{\text{d}} \) are the reference pressure (1000 hPa) and the partial pressure of dry air, \( R_d \) and \( R_v \) are the gas constants of dry air and water vapor, \( C_{pd} \) and \( C_{cl} \) are the heat capacities for dry air and liquid water, \( r \) and \( r_t \) are the water vapor mixing ratio and total water mixing ratio (although no condensed atmospheric water is considered here), \( H \) is relative humidity, and \( L_v \) is the latent heat of vaporization. While \( \theta_{eb} \) is a nonlinear function of temperature and specific humidity and should thus be calculated from data sampled on subdaily time scales, preliminary analysis (not shown) revealed extremely similar results, for this analysis of interannual variability, using either the monthly mean of four-times daily \( \theta_{eb} \), or the monthly mean of daily maximum \( \theta_{eb} \) [as in Nie et al. (2010)]. Additionally, we recognize that \( \theta_{eb} \) is not directly constrained by data in all regions and times represented in these reanalyses, but we do expect the assimilation of surface meteorological data, radiosondes, and SST to provide constraints in many regions. We are also somewhat reassured by the fact that two reanalysis datasets and two precipitation products yield qualitatively similar results in many regions, and by the fact that the summer mean structure of reanalyzed \( \theta_{eb} \) in South Asia was validated against radiosonde data by Boos and Kuang (2010).

For SST, we use the National Oceanic and Atmospheric Administration (NOAA) National Climatic Data Center (NCDC) extended reconstructed sea-surface temperature (ERSST2) database (Smith and Reynolds 2004). \( T_u \) and 700 hPa wind were obtained from the ERA-40 Reanalysis (Uppala et al. 2005).
Time series of regional precipitation anomalies were calculated for six monsoon regions using data from both the Global Historical Climatology Network [GHCN; (Peterson and Vose 1997)] and the Global Precipitation Climatology Project [GPCP; (Adler et al. 2003)]. The GPCP data set incorporates satellite information, includes precipitation over oceans, and is available from 1979 onwards. The GHCN data is derived from rain gauges, is available from 1900 on, includes data only over land, and is available in a globally gridded format from the University of Delaware’s Center for Climate Research (Matsuura and Willmott 2009).

Precipitation was averaged over six geographic areas chosen to reflect the interannual variability of regional monsoons and to generally be consistent with prior studies (Fig. 1). For instance, the West African domain (20°W-10°E; 10°N-20°N) corresponds to a typical area used for indices of Sahel rainfall (Janowiak 1988). The South Asian domain (60°E-100°E; 5°N-27°N) combines the prominent rainfall regions of the Indian summer monsoon and the Bay of Bengal (Wang and Fan 1999). The North American monsoon domain (115°W-100°E; 20°N-35°N) includes the regions of Mexico and the southwestern United States dominated by early and late monsoon season precipitation (Douglas et al. 1993; Gutzler 2004). The South American monsoon domain (70°W-40°W; 16°S-2°S) includes the region of maximum summer precipitation over central Brazil (Gan et al. 2004) and extends both west and east across the Amazon basin, consistent with the extent of the South Atlantic convergence zone (Vera et al. 2006). The East African domain (16°E-50°E; 24°S-9°S), although often not identified with a classical monsoon circulation, is the site of a seasonal tropical convergence zone associated with local maxima in $\theta_e$ and $T_u$ (McHugh 2004). The Australian domain (120°E-146°E; 20°S-5°S) is similar to that used for numerous indices of Australian monsoon rainfall (Colman et al. 2011; Wang and Ding 2008). Some of the regions chosen for averaging precipitation include the
climatological mean precipitation maximum while others are located on the periphery of that maximum. While this is not ideal, we chose to use rainfall indices similar to those in previous studies to facilitate comparison between our results and prior analysis of covariance with SST. There is qualitative agreement between the monsoon domains in Fig. 1 and the continental monsoon regions of Wang and Ding (2008).

We recognize that the dynamics responsible for precipitation may vary across these six regions, and that some of them may not be considered sites of classical monsoon circulations by some investigators, but we set those issues aside for now. Nie et al. (2010) found at least some consistency with a QE view of solstitial meridional flow in all of these regions, and our goal is simply to examine the interannual variability of precipitation in a similar QE framework.

2.2 Analysis methods

In this study we calculate correlation and regression coefficients for the following fields projected onto precipitation: SST, $\theta_{eb}$, $q_b$, $T_b$, 700 hPa horizontal wind, and $T_u$. All sub-cloud variables were evaluated on sigma levels about 20 hPa above the surface (as described above), and $T_u$ was averaged from 200 to 400 hPa. All analyses used three-month averaged time series for the full calendar year, except for the analyses of auto- and cross-correlations presented in Section 3.5, which used monthly data. Three-month averages were taken over the standard seasonal periods of Dec.-Feb., March-May, June-Aug., and Sept.-Nov., and only minor sensitivity to the choice of these periods was found. Use of three-month averages reduces the signal from subseasonal variability, and use of data for the entire year allows the influence of changes in the length of the rainy season to be included in our results. Using data for the entire year also allows possible changes in winter precipitation unrelated to summer monsoon
dynamics to influence the results, but preliminary analyses (discussed in the Appendix) show qualitatively similar results for full-year and local summer periods. Furthermore, since climatological monsoon onset and withdrawal dates can vary greatly between regions within the same hemisphere, use of a single “summer” period long enough to include all possible onset and withdrawal dates in each hemisphere would still allow some changes in dry season precipitation to influence results, but this influence would be greater in some regions than in others. We therefore opted for the simpler approach of using full-year data for every region. Having results that are relevant to annual-mean precipitation rather than just its local summer component might also be seen as an advantage for some applications. Details of these methodological decisions are outlined in the Appendix.

The focus of this paper is on analyses where we remove the linear variability associated with an index of ENSO. While study of the variations of \( \theta_{eb} \) related to ENSO might help in understanding the detailed mechanisms by which monsoon precipitation interacts with that climate signal, we leave that task for future work and here attempt to focus on less widely studied mechanisms that might be unrelated to ENSO. This is a simplistic approach and we acknowledge that our method does not wholly remove the ENSO signal (Compo and Sardeshmukh 2010). As a proxy for ENSO, we use the Nino 3.4 index (NOAA 2011), which is the monthly time series of SST anomalies from 5°S to 5°N and 170°W to 120°W. Variability linearly related to this index is removed using

\[
Y_{ij} = X_{ij} - \beta_{ij} N_{3.4},
\]

where \( X_{ij} \) is the original observed variable (e.g. \( \theta_{eb} \)) at a particular longitude \((i)\) and latitude \((j)\), and \( \beta_{ij} \) is the regression coefficient of \( X_{ij} \) projected onto the Nino 3.4 index \((N_{3.4})\). Specifically, \( X_{ij} \) is the linearly detrended time series of three-month average anomalies relative to the annual
cycle. The annual cycle was obtained by averaging monthly values over all years of the record.

$Y_{i,j}$ is the variable of interest following removal of the linear variability associated with the Nino 3.4 index, and we regress $Y_{i,j}$ onto the six regional precipitation indices, also linearly detrended, to identify spatial patterns that are linearly independent of that index.

Assessing statistical significance is complicated by the fact that the data possess spatial covariance and temporal autocorrelations. Autocorrelations of monsoon domain precipitation indices (spatially averaged over the boxes shown in Fig. 1) generally dropped below the 95% significance level within two months, and autocorrelations of regionally averaged $\theta_{eb}$ dropped below this level after a few months (see discussion of correlation functions in Section 3.5). Brown and Bretherton (1997) found monthly values of $\theta_{eb}$ and $T_u$ to have statistically significant autocorrelations at the 95% level for a representative time of four months and so reduced the number of degrees of freedom in their monthly data by a factor of four; we take the somewhat arbitrary but even more conservative approach of reducing our degrees of freedom by the same factor of four for three-monthly data. The decorrelation time scale for the three-monthly averaged precipitation time series is less than one three-month interval: within three months the autocorrelation function falls below the 95% confidence level. This makes correlation coefficients significant at the 95% level for a value of 0.3 for GHCN data and 0.4 for the shorter GPCP data. We use these thresholds for correlation coefficients to delineate regions with statistically significant linear relationships.

To evaluate antecedent relationships and persistence of precipitation and $\theta_{eb}$, sample auto- and cross-correlation functions were computed using the method of Box et al. (1994), with the lag-zero autocorrelation normalized to unity.
3 RESULTS

3.1 $\theta_{eb}$ vs. Precipitation

Regression coefficients for the three-month averaged, full-year time series of ERA-40 $\theta_{eb}$ anomalies projected onto GHCN precipitation anomalies are shown in Fig. 2 for the six monsoon regions. These results are based on data for 1958-2002, the period of the ERA-40 data, and the signal linearly related to the Nino 3.4 index was removed from $\theta_{eb}$ prior to the regression. Qualitatively similar results were obtained using local summer time series (see the Appendix for the example of West Africa).

For all six monsoon regions, positive precipitation anomalies are associated with positive continental $\theta_{eb}$ near and slightly poleward of the peak climatological precipitation. Positive regression patterns also occur over the nearby oceans for the South Asian, South American, and Eastern African monsoons.

In particular, enhanced Sahel precipitation is associated with a band of enhanced $\theta_{eb}$ that stretches across Africa at about 20°N (Fig. 2a). Enhanced South Asian monsoon precipitation is linearly related to enhanced $\theta_{eb}$ over northern India and the Himalayas, which are poleward of the box used for the precipitation index, as well as with enhanced $\theta_{eb}$ within and immediately south of that box (Fig. 2b). This is more spatially extensive than the area of positive correlations seen for the other monsoon regions, and the magnitudes of the regression coefficients are consistent with an increase in the land-ocean $\theta_{eb}$ gradient during strong years of the South Asian monsoon. Precipitation in the North American monsoon is positively related to $\theta_{eb}$ over Mexico and the southwestern United States, with the positive regression pattern lying poleward of the peak precipitation even though it is located squarely within the box used to define the precipitation index (Fig. 2c). Australian precipitation is positively related to $\theta_{eb}$ over much of
the Australian continent (Fig. 2d). Enhanced East African precipitation is associated with a band of positive $\theta_{eb}$ anomalies stretching from southern Africa to the southwestern Indian Ocean (Fig. 2e). South American monsoon precipitation is positively related to $\theta_{eb}$ in two regions, one on the eastern and poleward edge of the monsoon domain and extending over the Atlantic Ocean, and one poleward of the Amazon basin over Argentina. South American precipitation is also negatively related to $\theta_{eb}$ in the equatorial eastern Pacific and the northern tropical Atlantic (Fig. 2f).

We now present an alternate analysis of the same data, regressing limited zonal means of ERA-40 $\theta_{eb}$ onto the various GHCN monsoon precipitation indices, with the zonal means taken over the same longitude range used for averaging precipitation. This allows for the results to be interpreted in a single spatial dimension. Fig. 3 shows the summer mean (JJA and DJF) climatology of $\theta_{eb}$ as dashed lines, and the regression coefficients, scaled by a factor of $5 \text{ cm mo}^{-1}$ ($10 \text{ cm mo}^{-1}$ for Australia) and added to the $\theta_{eb}$ climatology, as solid lines. The gray envelopes around the solid lines indicate the 95% confidence intervals associated with the scaled regression coefficients. The scaling factors correspond approximately to the maximum amplitude of the precipitation anomalies, so that the solid lines represent $\theta_{eb}$ during wet monsoon years. For the South Asian monsoon, the $\theta_{eb}$ maximum simply increases in amplitude (Fig. 3b) during wet years, consistent with the hypothesis of Eltahir and Gong (1996). For the West African monsoon, the $\theta_{eb}$ maximum increases in amplitude and shifts poleward (Fig. 3a). A pattern common for all of the monsoon domains (except South Asia) is an increase of $\theta_{eb}$ confined to regions poleward of the summer $\theta_{eb}$ maxima (Figs. 3a, c, d, e, and f), which is distinct from the previously proposed idea that the meridional gradient of $\theta_{eb}$ increases equatorward of the summer maximum during wet years (Eltahir and Gong 1996). Although the meridional gradient
of $\theta_{eb}$ in the summer climatology is strong compared to the change in $\theta_{eb}$ between wet and dry years, the increase in $\theta_{eb}$ poleward of the summer maxima is statistically significant, is seen in five of the six monsoon regions, and in some regions extends well into the subtropics. South Asia also shows an increase in $\theta_{eb}$ poleward of the climatological $\theta_{eb}$ maximum during wet years, but unlike in the other regions this is part of a simple increase in the amplitude of the climatological maximum.

For all six regions, the regression patterns extend beyond the region within which the precipitation index was defined. However, the regression coefficients in areas remote from the monsoon regions (i.e. outside the regions shown in Figs. 2-7) were not statistically significant in some combinations of datasets examined in this study. For this reason, we focus here on the regression patterns local to the monsoon regions, and leave examination of the remote patterns, their sensitivity to data product, and their field significance for future work.

Although we further discuss possible mechanisms underlying the regression patterns in a later part of this paper, we note here that even a local positive relationship between $\theta_{eb}$ and precipitation is nontrivial because $\theta_{eb}$ is a conserved variable: $\theta_{eb}$ would not change if precipitation simply cooled and moistened low-level air without altering its specific entropy. A strong positive relationship between column water vapor and precipitation has been the subject of much investigation [e.g. Bretherton et al. (2004)], but those studies have been largely limited to oceanic regions and cannot easily explain the nature of continental patterns presented here.

### 3.2 Sensitivity to datasets

Analyses presented thus far used the globally gridded GHCN data set, which is based on continental rain gauge information. To test the sensitivity of our results to data source, we also
use the GPCP precipitation product, which has global coverage and is based on both satellite and rain gauge data. Regression analyses of $\theta_{eb}$, still calculated from the ERA-40 data, projected onto GPCP monsoon precipitation are presented in Fig. 4. These analyses used data from 1979-2002 (compared with 1958-2002 for Fig. 2), as the GPCP data begins in 1979 and the ERA-40 data ends at 2002.

The patterns of regression coefficients for the GPCP data are similar to those for the GHCN data in that most GPCP precipitation indices show a positive relationship with $\theta_{eb}$ near and just poleward of the peak summer mean precipitation. However, there are notable differences between the GHCN and GPCP patterns. The size of the statistically significant region has decreased for South Asia, and the largest regression coefficients are located closer to the equator (over southern India) in the GPCP data. Both the GPCP and GHCN regressions show a positive relationship between Sahel rainfall and $\theta_{eb}$ over large parts of northern Africa, but the region of positive regression coefficients is located further east, extends further north, and has higher peak amplitudes for the GPCP data. The statistically significant region has become almost vanishingly small in South America for the GPCP data, and the region of positive regression coefficients for East Africa has decreased greatly in size and is no longer located poleward of the mean precipitation peak. Patterns of ERA-40 $\theta_{eb}$ regressed onto GHCN precipitation for 1979-2002 (not shown) have more in common with Fig. 4 than with Fig. 2, while patterns for the earlier period 1958-1978 (also not shown) have more in common with Fig. 2 than with Fig. 4. These results suggest that the differences between Figs. 2 and 4 may reflect differences in interannual variability between the two time periods. The GPCP precipitation includes data over oceans, whereas the GHCN is limited to land, and this may also account for differences between Figs. 2 and 4.
To test the sensitivity of our results to the choice of reanalysis product, we also calculated \( \theta_{eb} \) from the NCEP Reanalysis (Fig. 5). As with the ERA-40 results in Fig. 4, we regress the NCEP \( \theta_{eb} \) onto the GPCP precipitation indices for 1979-2002. While there are differences in the amplitude and spatial extent of extrema, the patterns are quite similar to those obtained using ERA-40. In particular, most regions (but not South America or East Africa) show a positive relationship between precipitation and \( \theta_{eb} \) local to and immediately poleward of the mean precipitation peak.

3.3 \( q_b, T_b, \) and \( T_u \) vs. Precipitation

While variations of \( \theta_{eb} \) are of interest in their own right, it may be useful to know whether changes in humidity or temperature make a dominant contribution to the relationship between \( \theta_{eb} \) and precipitation. The positive relationship between precipitation and \( \theta_{eb} \) local to and slightly poleward of the index region is due mostly to the relationship between \( q_b \) and precipitation (Fig. 6). In contrast, regression patterns between \( T_b \) and monsoon region precipitation are, at least over continents, opposite in sign to those for \( \theta_{eb} \) (Fig. 7). Over oceans the relationship between \( T_b \) and precipitation typically has the same sign as that between \( \theta_{eb} \) and precipitation. For instance, the region of high \( \theta_{eb} \) over the southern Indian Ocean that is associated with enhanced East African precipitation has the same sign for regressions of \( \theta_{eb}, q_b, \) and \( T_b \), but the regression coefficient over continental Africa is negative for \( T_b \) and positive for both \( \theta_{eb} \) and \( q_b \). This is consistent with a simple scenario in which the oceanic patterns result from SST changes that control the thermodynamic state of the subcloud layer without changes in subcloud relative humidity or air-sea temperature difference. Explaining these patterns over land is less straightforward, because while one does expect precipitation over land to be accompanied by
enhanced specific humidity and reduced surface air temperature (Eltahir 1998; Trenberth and Shea 2005), changes in $\theta_{eb}$ signify a net change in the entropy of the subcloud layer rather than just its partitioning between sensible and latent components. Eltahir (1998) hypothesized that soil moisture could take part in a local positive feedback on precipitation by enhancing subcloud energy content through reductions in both surface albedo and surface longwave emissions. The local positive relationships between $\theta_{eb}$ and precipitation seen here are consistent with that hypothesis, although interactions with large-scale dynamics may also operate. Further investigation would be necessary to establish the influence that soil moisture has on $\theta_{eb}$, and to distinguish it from other influences such as large-scale advection of $\theta_{eb}$ or other controls on surface fluxes.

The statistically significant patterns for $q_b$ and $T_b$ regressed on South Asian precipitation are less spatially extensive than the regression patterns for $\theta_{eb}$. This results from the nonlinear dependence of $\theta_{eb}$ on $q_b$ (see Appendix) and illustrates the power of using $\theta_{eb}$ instead of surface air temperature or humidity to interpret variations in monsoon strength.

Also plotted in Fig. 6 (as green contours) are correlation coefficients between precipitation and $T_u$. Consistent with a QE framework, maxima of $\theta_{eb}$ and $T_u$ are spatially coincident in many of Earth’s regional summer monsoons (Nie et al. 2010), so one might expect interannual variations of those quantities to have similar patterns if QE holds. Indeed, $T_u$ does seem to covary with precipitation in a somewhat similar fashion to $\theta_{eb}$ in some regions. The correlations between monsoon precipitation and $T_u$ are neither as pronounced as, nor exactly collocated with, the peak correlations between precipitation and $\theta_{eb}$. However, there are similarities worth noting. For example, positive correlations are seen between Australian precipitation and $T_u$ over central Australia. The positive correlation of Sahel precipitation with $T_u$ over the eastern
Atlantic and Mediterranean Sea is located well poleward of the peak $\theta_{eb}$ signal and only has small regions that exceed the statistically significant threshold of 0.3. Nevertheless, the relationships of Sahel precipitation with $T_u$, $\theta_{eb}$, and $q_b$ may be consistent with the idea that positive anomalies of Mediterranean SST produce enhanced Sahel rainfall through advection of moisture across the Sahara (Rowell 2003). No strong, local correlation between precipitation and $T_u$ is seen in the Americas.

It is intriguing that the positive correlations between precipitation and $T_u$ in several regions have an equatorial minimum, with maxima in either hemisphere roughly equidistant from the equator. This meridional structure, symmetric about the equator, is seen for the Sahel, South Asia, and Australia (Figs. 6a, b, and d, respectively), though the winter hemisphere maxima lie outside of the areas plotted in Fig. 6 and are not shown here. This type of temperature structure is consistent with an angular momentum-conserving Hadley circulation that has an off-equatorial ascent branch (Lindzen and Hou 1988) and can be seen in simple models of monsoon flow which couple $\theta_{eb}$ with $T_u$ in a QE framework [e.g. Emanuel (1995) and Boos and Emanuel (2008b)].

Excitation of n=1 equatorial Rossby waves might also produce equatorially symmetric anomalies through entirely linear dynamics (Matsuno 1966). Both Gill (1980) and Rodwell and Hoskins (1996) show a stationary response both north and south of the equator, at locations west of idealized off-equatorial steady heating.

### 3.4 Wind vs. precipitation

While a thorough study of the covariation of precipitation with large-scale winds is beyond the scope of this manuscript, we do examine correlations between precipitation and flow in the lower free-troposphere. This is motivated by the relatively recent finding that interannual
variations of Sahel precipitation in climate models is related to the strength of the Sahara low
(Biasutti et al. 2009; Haarsma et al. 2005; Thorncroft et al. 2011) and by the identification of
shallow circulations in the seasonal mean climatologies of summer meridional flow in Australia
and southern Africa [e.g. Nie et al. (2010)]. These shallow circulations consist of near-surface
poleward flow that extends beyond the primary monsoon precipitation maximum and terminates
near the peak subcloud potential temperature maximum ($\theta_b$, not $\theta_{eb}$). Since the positive
anomalies of $\theta_{eb}$ and $q_b$ that are associated with enhanced monsoon precipitation also extend
poleward of the main precipitation maximum into the desert regions of Australia and West
Africa, it seems reasonable to ask if they are associated with some change in the shallow flow. A
working hypothesis is that the advection of dry air by the shallow circulation into the primary
monsoon convection zone reduces the rate of monsoon precipitation. This is consistent with
discussion of the possible influence of shallow flow in the Sahara low on the seasonal cycle of
Sahel precipitation by Zhang et al. (2008); Peyrillé and Lafore (2007) present some relevant
numerical model results.

Figure 8 presents regressions of both the full precipitation field (shading) and the 700 hPa
horizontal wind (magenta vectors) projected onto the monsoon domain precipitation indices.
Also plotted, for comparison, is the summer-mean (JJA and DJF) 700 hPa horizontal wind (gray
vectors). The 700 hPa level is above or near the top of the low-level, poleward branch of the
deep-baroclinic monsoon flow, and is within the upper, equatorward branch of the shallow flow
(which projects strongly onto a second-baroclinic mode). Thus, for those monsoons with a
pronounced shallow meridional circulation, such as West Africa and Australia, the 700 hPa mean
summer flow is directed toward the equator in the region of the precipitation maximum. This
equatorward flow is part of a broader 700 hPa anticyclone centered over the desert regions.
Equatorward flow is also seen in the southwest part of the box over which precipitation was averaged in East Africa, and northwest of India in an intense heat low (Ramage 1966).

Regression analysis shows that enhanced precipitation is associated with enhanced poleward flow at 700 hPa in most regions (especially in Australia, the Sahel, and East Africa), which could alternatively be viewed as a weakening of the climatological equatorward flow at that level. This weakening of the 700 hPa flow could be due to a weakening of the dry, shallow circulation or a vertical expansion of that circulation. It could also be due to a vertical expansion of the lower, poleward branch of the deep-baroclinic circulation or a strengthening of that first-baroclinic part of the flow, but such first-baroclinic changes would not be expected to produce the 700 hPa cyclones centered over the deserts that are seen in the regression patterns (e.g. Figs. 8d, e). Both Haarsma (2005) and Biasutti et al. (2009) found that enhanced Sahel precipitation was associated with negative anomalies of sea level pressure and 925 hPa geopotential height in the Sahara. This is inconsistent with a generic weakening of the shallow heat-low circulation during times of enhanced rainfall, and indicates that further investigation of the relevant dynamics is needed. The fact that subcloud temperatures over continental Australia are negatively correlated with Australian precipitation (Fig. 7d) would also seem inconsistent with a strengthening of the Australian heat low during times of enhanced monsoon precipitation. We plan to more thoroughly examine interannual variations in the intensity and vertical structure of shallow flow over both Australia and northern Africa in future work.

Even in regions without a well-defined shallow circulation, the change in 700 hPa flow during wet years is clearly distinct from a simple strengthening of the climatological mean flow at that level. For example, the regression vectors are roughly perpendicular to the mean flow vectors within the eastern half of the box used to define the precipitation index in South Asia,
indicating an association between positive precipitation anomalies and enhanced poleward flow at 700 hPa. In North America, enhanced monsoon precipitation is associated with anomalous southwesterly monsoon flow from the Pacific, in contrast to the mean flow from the Gulf of Mexico. In South America, enhanced monsoon precipitation corresponds, along the poleward edge of the monsoon domain, to a reduction in the strength of the northerly low-level jet at 700 hPa. This is consistent with the results of Garreaud and Wallace (1998), where they show the development of cold fronts and convection when cold and dry midlatitude air is advected into subtropical South America. These results also show that enhanced precipitation does not simply result from enhanced low-level moisture in the absence of circulation changes.

Regardless of the reasons for these changes in the 700 hPa flow, they are consistent with the hypothesis that stronger shallow circulations inhibit monsoon precipitation by advecting low \( \theta_{eb} \) (i.e. dry) air into the monsoon domain at low levels. This might be seen as a variation of the “ventilation” mechanism of Chou et al. (2001) and Chou and Neelin (2003) in which advection of low moist static energy air limited the poleward extent of a monsoon, but via a shallow circulation not represented in those models. The enhancement of \( \theta_{eb} \) poleward of the primary precipitation maximum also seems consistent with application of a ventilation mechanism to interannual variability, but our analyses do not determine the relative importance of variations in the strength of a shallow circulation and the variations in \( \theta_{eb} \) (which may interact with the precipitation maxima entirely through deep, first-baroclinic dynamics).

### 3.5 Antecedent Relationships

To explore lead-lag relationships, we completed correlation function analyses for the Australian region using the precipitation index and a \( \theta_{eb} \) index computed for a region that
includes continental Australia (120-150°E, 0-42°S). Correlation function estimates (Fig. 9) show that the autocorrelations of precipitation are significant only out to one month of lead time, and autocorrelations of $\theta_{eb}$ are significant for about two months. In contrast, correlations between $\theta_{eb}$ and precipitation are statistically significant, albeit weak, when $\theta_{eb}$ leads precipitation by five months, and the fact that cross-correlations at these lead times are greater than either of the autocorrelations is suggestive of an antecedent relationship due to more than simple persistence. The cross-correlation function falls rapidly below the significance threshold when precipitation leads $\theta_{eb}$. When precipitation lags by 1-5 months (left of the origin), the precipitation-$\theta_{eb}$ relationship is significant at the 95% confidence level and only slightly weaker than the precipitation-ENSO correlation. At zero-lag, the precipitation-$\theta_{eb}$ relationship is at least as strong as the precipitation-ENSO relationship.

4 CONCLUSIONS

We have examined the covariation of seasonal precipitation anomalies in six monsoon regions with local $\theta_{eb}$ and several related variables. We chose to remove the variability linearly associated with the Nino-3.4 SST index in order to reduce the likelihood that our results will simply provide another perspective on the well-known association between ENSO and monsoon precipitation. We found positive precipitation in each region to be associated with enhanced $\theta_{eb}$ local to and slightly poleward of the primary monsoon precipitation maximum. This general local result held for both of the reanalysis products used to calculate $\theta_{eb}$. The continental variations in $\theta_{eb}$ local to the precipitation maxima are mostly due to variations in $q_b$, with local changes in $T_h$ having an opposite sign. $T_u$ was also shown to covary with $\theta_{eb}$ and precipitation in ways qualitatively consistent with a strict QE hypothesis in a few regions: during times of
enhanced monsoon precipitation, regions of enhanced $T_u$ are located over the regions of
enhanced $\theta_{eb}$ slightly poleward of the peak monsoon ascent zone in Australia and South Asia,
with weak signals in western and eastern Africa.

Here we summarize three distinct ways in which boundary layer moist entropy may covary
with monsoon precipitation, two of which are previously recognized relationships and a third that
is relevant to the results presented in this study. For the South Asian monsoon, the
climatological $\theta_{eb}$ peak simply increases in amplitude during years of enhanced precipitation.
This is illustrated schematically in Fig. 10a, can also be seen in Fig. 3b, and is consistent with
simple theoretical ideas for how monsoon strength relates to the amplitude of an off-equatorial
$\theta_{eb}$ maximum [e.g. Emanuel (1995), Prive and Plumb (2007a)]. Eltahir and Gong (1996) argued
that for West Africa during wet years, the meridional gradient of boundary layer moist entropy
may increase due to a reduction in boundary layer moist entropy on the winter side of the
equator, as shown schematically in Fig. 10b. Thus, panels a and b in the schematic of Fig. 10
depict previously recognized relationships by which monsoon precipitation is enhanced when the
meridional gradient of $\theta_{eb}$ increases between the $\theta_{eb}$ peak and the winter hemisphere. We
recognize here an additional relation between monsoon precipitation and $\theta_{eb}$, whereby wet
monsoon years are associated with enhanced $\theta_{eb}$ poleward of the climatological $\theta_{eb}$ maxima (see
schematic in Fig. 10c, and also Figs. 3a, c-f).

While most theoretical studies of monsoon dynamics that used a QE framework employ
only a single baroclinic mode [e.g. Emanuel (1995) and Chou et al. (2001)], recent work has
suggested that dry, shallow circulations with low-level temperature maxima poleward of the
region of peak monsoon precipitation may play a role in the mean state and interannual
variability of monsoons [e.g. Haarsma et al. (2005), Nie et al. (2010), Biasutti et al. (2009)]. Our
results show that the low-level moisture anomalies associated with variations in monsoon precipitation may extend into the desert regions which serve as thermal maxima for these shallow circulations, and that the climatological equatorward flow at 700 hPa weakens during seasons of enhanced monsoon precipitation. This is illustrated schematically in Fig. 11, though we emphasize that it is unclear whether the increase in precipitation is due to a weakening or deepening of the shallow circulation, or an increase in its moisture content. In any case, moist convection has been shown to be sensitive to the presence of dry air in the lower free-troposphere (Derbyshire et al. 2004; Raymond and Zeng 2000; Tompkins 2001), and these findings suggest a particular application of this idea to interannual monsoon variability. An alternate possibility is that the desert regions and their associated shallow circulations are being modulated by Rossby waves emitted by deep cumulus convection (Bollasina and Nigam 2011; Rodwell and Hoskins 1996), with the crucial dynamics involving only deep, first-baroclinic motions. Regardless of the mechanism, our results emphasize the potential importance of low-level moisture anomalies poleward of the precipitation peak for the interannual variability of monsoons.

The covariation of deserts with monsoon precipitation might be evidence for a variation on the hypothesis of Charney (1975), in which monsoon precipitation interacts with the surface albedo of an adjacent desert. The positive anomalies of $\theta_{eb}$ over deserts seen in our analyses could be caused by negative anomalies of land surface albedo, which in turn could result from changes in the latitude of the vegetation-desert boundary or simply from the darkening of wet soil. A shift in the latitude of the desert boundary is included in the schematic of Fig. 11, and although the existence of such a shift is speculative, it would be consistent with the fact that precipitation anomalies extend into the equatorward edge of desert regions (Fig. 8).
The lack of a clear association between South Asian precipitation and $\theta_e$ poleward of that region’s climatological $\theta_e$ maximum is consistent with the idea that the Himalayas and the Hindu Kush mountains insulate the South Asian monsoon from the influence of nearby deserts (Boos and Kuang 2010; Boos and Hurley 2013; Chakraborty et al. 2006). If topography inhibits variations in the properties of Asian deserts from influencing the South Asian monsoon, one would expect interannual variability in that region to differ from that of monsoons which are coupled with variations in adjacent deserts.

The linear analyses presented here are only intended to be a first step in examination of the interannual variability of observed monsoons in a QE framework. Our discussion of mechanisms was almost entirely speculative, and needs to be assessed in quantitative theoretical frameworks. It is unclear whether existing QE models such as the Quasi-equilibrium Tropical Circulation Model [QTCM, Neelin and Zeng (2000)] or the zonally symmetric model used by Boos and Emanuel (2008a, 2008b) are suitable for such a task, as their dynamical structure cannot represent shallow heat-low flow. It is also unclear how well dry, shallow circulations are represented in more comprehensive global climate models. But regardless of theoretical progress, further examination of the observed statistics of subcloud thermodynamics may help in understanding interannual variations in monsoons.
APPENDIX

In this Appendix we present five figures, both for reference and to illustrate the effects of some of the methodological choices.

Summer mean $\theta_{eb}$ climatologies are presented in Fig. A1, for comparison with the regression patterns shown in the body of this paper. These climatologies show the locations of the summer mean $\theta_{eb}$ maxima, and are consistent with those shown in Nie et al. (2010).

Over ocean, correlations between monsoon precipitation and SST are expected to be highly similar to correlations between precipitation and $\theta_{eb}$ if the air-sea temperature difference and subcloud relative humidity are fixed. Figure A2 shows correlations between West African precipitation and both SST and $\theta_{eb}$, and is broadly consistent with a scenario in which variations in $\theta_{eb}$ over ocean have, at least, the same sign as variations in SST. For Fig. A2, the variability in $\theta_{eb}$ linearly associated with ENSO (which is shown in Fig. A3) has not been removed and only local summer (June-Aug.) data was used, for easier comparison with previous work. The positive signal in the north Atlantic and the negative signals in the extratropical south Atlantic and western Indian Oceans (Fig. A2) qualitatively agree with other published correlations between West African monsoon precipitation and SST (Caminade and Terray 2010; Lu 2009). Of course, $\theta_{eb}$ is not a function of only SST; even when a 6-month low-pass filter was used, Brown and Bretherton (1997) found a correlation coefficient of only 0.60 between SST and $\theta_{eb}$, with variations in oceanic $\theta_{eb}$ being due almost entirely to variations in $q_b$. They also found that $T_u$ was correlated more strongly with $\theta_{eb}$ than with SST, which provides strong motivation for the study of variations in $\theta_{eb}$ in general. Another obvious advantage of examining $\theta_{eb}$ is that it extends the analysis domain over land without, apparently, introducing dramatic changes to previously studied relationships in which SST was assumed to strongly control near-surface
thermodynamics (e.g. Fig. A2).

Figure A4 shows results from an analysis similar to that of Fig. A2, but with regression coefficients plotted instead of correlation coefficients, and with removal of the variability linearly associated with Nino-3.4. Patterns that are statistically significant at the 95% confidence level are outlined in thick black contours. Removal of the ENSO signal eliminates the strong, statistically significant negative relationship in the eastern tropical Pacific (not shown) but leaves a strong positive relationship in the North Atlantic between precipitation and both SST and $\theta_{eb}$. The positive relationship with $\theta_{eb}$ extends in a zonally elongated band across the northern part of the Sahel. These sorts of relationships between precipitation and continental $\theta_{eb}$ were examined in the Results section, with Fig. A4 presented here to illustrate the effects of removing Nino-3.4 variability and of performing analyses for local summer instead of the full calendar year (compare with Fig. 2a).

The regression analyses for South Asian precipitation produced patterns of regression coefficients that were more spatially extensive for $\theta_{eb}$ than for $q_b$ or $T_b$ (compare Figs. 2b, 6b, and 7b). To show why, we calculate $\theta_{eb}$ allowing only $q_b$ or $T_b$ to vary, and then repeat the regression analysis (subcloud pressure is held constant in both of these alternate calculations of $\theta_{eb}$). When only $q_b$ is allowed to vary in the calculation of $\theta_{eb}$, the pattern of regression coefficients closely matches that computed using the true $\theta_{eb}$ field (compare Figs. A5a and 2b). This indicates that the exponential dependence of $\theta_e$ on $q$ can account for the differences between Figs. 2b and 6b. When only $T_b$ varies in the calculation of $\theta_{eb}$, the regression pattern closely resembles that of observed $T_b$ onto precipitation (compare Figs. A5b and 7b), consistent with the linear relationship between $\theta_e$ and $T$. 
Acknowledgements: ERA-40 and NCEP reanalyses were obtained from the Research Data Archive (RDA) which is maintained by the Computational and Information Systems Laboratory (CISL) at the National Center for Atmospheric Research (NCAR). NCAR is sponsored by the National Science Foundation (NSF). The original data are available from the RDA (http://dss.ucar.edu) in dataset numbers ds090.0 (NCEP) and ds126.2 (ERA-40). We used a netcdf format version of the GHCN data available from the University of Washington’s Joint Institute for the Study of the Atmosphere and Ocean (JISAO 2009). Criticism by three anonymous reviewers improved the manuscript. Zhiming Kuang and Adam Sobel made helpful comments. We gratefully acknowledge financial support from Office of Naval Research YIP award N00014-11-1-0617 and from Yale University.
REFERENCES


31


JISAO: University of Delaware global surface air temperature and precipitation climatology and monthly grids, 1900-2008.


NOAA, 2011: Nino 3.4 SST anomalies. C. P. Center, Ed.


Fig. 1. Local summer mean precipitation, (a) June-July-August (JJA) mean with boxes drawn around the regions used to define precipitation indices in West Africa, South Asia, and North America, and (b) December-January-February (DJF) with boxes indicating precipitation index regions for Australia, East Africa, and South America. Precipitation data is the Global Historical Climatology Network (GHCN, shading, [cm mo\(^{-1}\)]) and the Global Precipitation Climatology Project (GPCP, solid contours, contour interval of 5 cm mo\(^{-1}\)).

Fig. 2. Regression coefficients (shading, [K][cm mo\(^{-1}\)]\(^{-1}\)) of three-month averaged full year \(\theta_{eb}\) (ERA-40) onto monsoon region precipitation (GHCN), for 1958-2002, following removal of the linear variability associated with ENSO, for (a) West Africa, (b) South Asia, (c) North America, (d) Australia, (e) East Africa, and (f) South America. Precipitation was averaged over the boxed regions. Black contours (negative dashed) delineate regions where the correlation coefficient is statistically significant at the 95% confidence level. Note the different color scale for panel a.

Fig. 3. Dashed lines show limited zonal means of local summer (JJA and DJF) mean ERA-40 \(\theta_{eb}\). Solid lines show \(\theta_{eb}\) during a year of high precipitation, constructed by adding the summer mean \(\theta_{eb}\) to the regression coefficient scaled by a precipitation anomaly of 5 cm mo\(^{-1}\) (10 cm mo\(^{-1}\) for Australia). Regression coefficients were calculated using GHCN precipitation and limited zonal means of \(\theta_{eb}\). Panels show results for a) West Africa, b) South Asia, c) North America, d) Australia, e) East Africa, and f) South America. Limited zonal means were computed across the longitudes of the boxes in Fig. 2.
Fig. 4. As in Fig. 2, but for regression of ERA-40 $\theta_{eb}$ onto GPCP monsoon region precipitation for 1979-2002, with regression coefficients (shading) in units of [K][cm mo$^{-1}$]$^{-1}$.

Fig. 5. As in Fig. 2, but for regression of NCEP $\theta_{eb}$ onto GPCP monsoon region precipitation for 1979-2002, with regression coefficients (shading) in units of [K][cm mo$^{-1}$]$^{-1}$.

Fig. 6. Regression coefficients (shading, [g kg$^{-1}$][cm mo$^{-1}$]$^{-1}$) of three-month averaged full year $q_b$ (ERA-40) onto monsoon region precipitation (GHCN), for 1958-2002, following removal of the linear variability associated with ENSO, for (a) West Africa, (b) South Asia, (c) North America, (d) Australia, (e) East Africa, and (f) South America. Green contours (0.10 contour interval, zero contour not shown) are correlation coefficients between precipitation (GHCN) and upper troposphere temperatures ($T_u$, ERA-40; 200 - 400 hPa), dashed contours indicate negative correlations (values are significant at values above or below 0.3 or -0.3, bold contour).

Precipitation was averaged over the boxed regions. Black contours delineate regions where the correlation coefficient between $q_b$ and precipitation is statistically significant at the 95% confidence level.

Fig. 7. Regression coefficients (shading, [K][cm mo$^{-1}$]$^{-1}$) of three-month averaged full year $T_b$ (ERA-40) onto monsoon region precipitation (GHCN), for 1958-2002, following removal of the linear variability associated with ENSO, for (a) West Africa, (b) South Asia, (c) North America, (d) Australia, (e) East Africa, and (f) South America. Precipitation was averaged over the boxed regions. Black contours delineate regions where the correlation coefficient is statistically significant at the 95% confidence level.
Fig. 8. Regression coefficients (shading, \([\text{cm mo}^{-1}] [\text{cm mo}^{-1}] \cdot \text{s}^{-1}\)) of three-month averaged full year precipitation (GHCN) onto monsoon region precipitation, for 1958-2002, following removal of the linear variability associated with ENSO, for (a) West Africa, (b) South Asia, (c) North America, (d) Australia, (e) East Africa, and (f) South America. Vectors are horizontal winds at 700hPa. Gray vectors are the local summer climatological winds \([\text{m s}^{-1}]\) and magenta vectors are the coefficients \([\text{m s}^{-1}] [\text{cm mo}^{-1}] \cdot \text{s}^{-1}\) of winds regressed onto precipitation. Reference vectors (inset boxes) are scaled to the value indicated, in units of \([\text{m s}^{-1}]\) for the gray vectors and units of \([\text{m s}^{-1}] [\text{cm mo}^{-1}] \cdot \text{s}^{-1}\) for the magenta vectors. Regression shading and vectors are plotted only where the regression of precipitation onto the precipitation index or either the zonal or meridional wind onto the precipitation index is statistically significant at the 95% confidence level. Precipitation was averaged over the boxed regions.

Fig. 9. Australian monsoon auto- and cross-correlation functions, for 1979-2002. GPCP precipitation was averaged from 120°E-146°E, 5°S-20°S and ERA-40 \(\theta_{eb}\) from 120°E-150°E, 0°S-42°S. Auto-correlations are for precipitation (dashed gray line) and \(\theta_{eb}\) (solid gray line). Cross-correlations are for precipitation versus \(\theta_{eb}\) (solid black line) and precipitation versus the Nino 3.4 index (dotted gray line). Negative lag indicates that \(\theta_{eb}\) or the Nino 3.4 index leads precipitation. Horizontal gray bar indicates correlation values which are insignificant at the 95% confidence level.

Fig. 10. Schematic of ways in which monsoon precipitation covaries with \(\theta_{eb}\). Solid gray lines represent the basic state and the dashed black line shows \(\theta_{eb}\) during years of high precipitation.
Gray ovals indicate latitudes where positive or negative regression patterns would occur for $\theta_{eb}$ regressed on precipitation indices. The meridional gradient of $\theta_{eb}$ on the equatorward side of the $\theta_{eb}$ peak increases during wet years in (a) and (b), both of which were illustrated schematically by Eltahir and Gong (1996, their Figs. 9 and 13, respectively). An increase of $\theta_{eb}$ poleward of the basic state $\theta_{eb}$ maximum is shown in (c), with negligible change at the location of the basic state maximum.

Fig. 11. Schematic of the relationships seen in our regression analyses, for years of a) normal precipitation and b) high precipitation. Diagrams depict the shallow meridional circulation (dashed black overturning cell) associated with a desert heat low, the deep monsoon circulation (solid gray overturning cell), and the $\theta_{eb}$ profile (dashed black line is for a wet year and solid gray line is for the basic state, repeated in (b) for reference). A poleward contraction of the desert during wet years is included in (b), based only on speculation.

Fig. A1. JJA (a) and DJF (b) mean $\theta_{eb}$ (ERA-40, 1958-2002). Contour interval, 5K. Boxes indicate monsoon domains used for precipitation indices.

Fig. A2. Correlation coefficients for JJA, 1979-2002, for West African monsoon precipitation (GHCN) and (a) SSTs (ERSST) and (b) $\theta_{eb}$ (ERA-40). Precipitation was averaged over the boxed region.
Fig. A3. Regression coefficients of three-month averaged full-year seasonal anomalies of $\theta_{eb}$ (ERA-40, 1958-2002) onto the Nino 3.4 index. The zero contour is not drawn, the contour interval is 0.5 K/K, solid lines are positive, and dashed lines are negative.

Fig. A4. Regression coefficients (shading, [K][cm mo]\(^{-1}\)) for JJA, 1958-2002, for (a) SSTs (ERSST) and (b) $\theta_{eb}$ (ERA-40) onto West African monsoon precipitation (GHCN), following removal of the linear variability associated with ENSO. Precipitation was averaged over the boxed region. Black contours delineate regions where the correlation coefficient is statistically significant at the 95% confidence level. Note the different color scales.

Fig. A5. Regression coefficients of modified versions of $\theta_{eb}$ (ERA-40) onto South Asian monsoon precipitation (GHCN). Color shading and black lines represent the regression coefficients ([K][cm mo]\(^{-1}\)) and the 95% confidence level as in Fig. 2. Prior to regression analysis, $\theta_{eb}$ is calculated (a) with $T_b$ and subcloud pressure held constant, or (b) with $q_b$ and subcloud pressure held constant.
Fig. 1. Local summer mean precipitation, (a) June-July-August (JJA) mean with boxes drawn around the regions used to define precipitation indices in West Africa, South Asia, and North America, and (b) December-January-February (DJF) with boxes indicating precipitation index regions for Australia, East Africa, and South America. Precipitation data is the Global Historical Climatology Network (GHCN, shading, [cm mo⁻¹]) and the Global Precipitation Climatology Project (GPCP, solid contours, contour interval of 5 cm mo⁻¹).
Fig. 2. Regression coefficients (shading, [K][cm mo$^{-1}$]$^{-1}$) of three-month averaged full year $\theta_{eb}$ (ERA-40) onto monsoon region precipitation (GHCN), for 1958-2002, following removal of the linear variability associated with ENSO, for (a) West Africa, (b) South Asia, (c) North America, (d) Australia, (e) East Africa, and (f) South America. Precipitation was averaged over the boxed regions. Black contours (negative dashed) delineate regions where the correlation coefficient is statistically significant at the 95% confidence level. Note the different color scale for panel (a).
Fig. 3. Dashed lines show limited zonal means of local summer (JJA and DJF) mean ERA-40 $\theta_{eb}$. Solid lines show $\theta_{eb}$ during a year of high precipitation, constructed by adding the summer mean $\theta_{eb}$ to the regression coefficient scaled by a precipitation anomaly of 5 cm mo$^{-1}$ (10 cm mo$^{-1}$ for Australia). Regression coefficients were calculated using GHCN precipitation and limited zonal means of $\theta_{eb}$. Panels show results for a) West Africa, b) South Asia, c) North America, d) Australia, e) East Africa, and f) South America. Limited zonal means were computed across the longitudes of the boxes in Fig. 2.
Fig. 4. As in Fig. 2, but for regression of ERA-40 $\theta_{eb}$ onto GPCP monsoon region precipitation for 1979-2002, with regression coefficients (shading) in units of $[K][cm \ mo^{-1}]^{-1}$. 
Fig. 5. As in Fig. 2, but for regression of NCEP $\theta_{eb}$ onto GPCP monsoon region precipitation for 1979-2002, with regression coefficients (shading) in units of [K/cm mo$^{-1}$]$^{-1}$. 
Fig. 6. Regression coefficients (shading, [g kg⁻¹][cm mo⁻¹]⁻¹) of three-month averaged full year $q_b$ (ERA-40) onto monsoon region precipitation (GHCN), for 1958-2002, following removal of the linear variability associated with ENSO, for (a) West Africa, (b) South Asia, (c) North America, (d) Australia, (e) East Africa, and (f) South America. Green contours (0.10 contour interval, zero contour not shown) are correlation coefficients between precipitation (GHCN) and upper troposphere temperatures ($T_u$, ERA-40; 200 - 400 hPa), dashed contours indicate negative correlations (values are significant at values above or below 0.3 or -0.3, bold contour). Precipitation was averaged over the boxed regions. Black contours delineate regions where the correlation coefficient between $q_b$ and precipitation is statistically significant at the 95% confidence level.
Fig. 7. Regression coefficients (shading, $[K][\text{cm mo}^{-1}]$) of three-month averaged full year $T_b$ (ERA-40) onto monsoon region precipitation (GHCN), for 1958-2002, following removal of the linear variability associated with ENSO, for (a) West Africa, (b) South Asia, (c) North America, (d) Australia, (e) East Africa, and (f) South America. Precipitation was averaged over the boxed regions. Black contours delineate regions where the correlation coefficient is statistically significant at the 95% confidence level.
Fig. 8. Regression coefficients (shading, [cm mo$^{-1}$][cm mo$^{-1}$]$^{-1}$) of three-month averaged full year precipitation (GHCN) onto monsoon region precipitation, for 1958-2002, following removal of the linear variability associated with ENSO, for (a) West Africa, (b) South Asia, (c) North America, (d) Australia, (e) East Africa, and (f) South America. Vectors are horizontal winds at 700hPa. Gray vectors are the local summer climatological winds [m s$^{-1}$] and magenta vectors are the coefficients [m s$^{-1}$][cm mo$^{-1}$]$^{-1}$ of winds regressed onto precipitation. Reference vectors (inset boxes, bottom right) are scaled to the value indicated, in units of [m s$^{-1}$] for the gray vectors and units of [m s$^{-1}$][cm mo$^{-1}$]$^{-1}$ for the magenta vectors. Regression shading and vectors are plotted only where the regression of precipitation onto the precipitation index or either the zonal or meridional wind onto the precipitation index is statistically significant at the 95% confidence level. Precipitation was averaged over the boxed regions.
Fig. 9. Australian monsoon auto- and cross-correlation functions, for 1979-2002. GPCP precipitation was averaged from 120°E-146°E, 5°S-20°S and ERA-40 $\theta_{eb}$ from 120°E-150°E, 0°S-42°S. Auto-correlations are for precipitation (dashed gray line) and $\theta_{eb}$ (solid gray line). Cross-correlations are for precipitation versus $\theta_{eb}$ (solid black line) and precipitation versus the Nino 3.4 index (dotted gray line). Negative lag indicates that $\theta_{eb}$ or the Nino 3.4 index leads precipitation. Horizontal gray bar indicates correlation values which are insignificant at the 95% confidence level.
Fig. 10. Schematic of ways in which monsoon precipitation covaries with $\theta_{eb}$. Solid gray lines represent the basic state and the dashed black line shows $\theta_{eb}$ during years of high precipitation. Gray ovals indicate latitudes where positive or negative regression patterns would occur for $\theta_{eb}$ regressed on precipitation indices. The meridional gradient of $\theta_{eb}$ on the equatorward side of the $\theta_{eb}$ peak increases during wet years in (a) and (b), both of which were illustrated schematically by Eltahir and Gong (1996, their Figs. 9 and 13, respectively). An increase of $\theta_{eb}$ poleward of the basic state $\theta_{eb}$ maximum is shown in (c), with negligible change at the location of the basic state maximum.
Fig. 11. Schematic of the relationships seen in our regression analyses, for years of a) normal precipitation and b) high precipitation. Diagrams depict the shallow meridional circulation (dashed black overturning cell) associated with a desert heat low, the deep monsoon circulation (solid gray overturning cell), and the $\theta_{eb}$ profile (dashed black line is for a wet year and solid gray line is for the basic state, repeated in (b) for reference). A poleward contraction of the desert during wet years is included in (b), based only on speculation.
Fig. A1. JJA (a) and DJF (b) mean $\theta_{eb}$ (ERA-40, 1958-2002). Contour interval, 5K. Boxes indicate monsoon domains used for precipitation indices.
Fig. A2. Correlation coefficients for JJA, 1979-2002, for West African monsoon precipitation (GHCN) and (a) SSTs (ERSST) and (b) \( \theta_{eb} \) (ERA-40). Precipitation was averaged over the boxed region.
Fig. A3. Regression coefficients of three-month averaged full-year seasonal anomalies of $\theta_{eb}$ (ERA-40, 1958-2002) onto the Nino 3.4 index. The zero contour is not drawn, the contour interval is 0.5 K/K, solid lines are positive, and dashed lines are negative.
Fig. A4. Regression coefficients (shading, [K][cm mo⁻¹]⁻¹) for JJA, 1958-2002, for (a) SSTs (ERSST) and (b) $\theta_{eb}$ (ERA-40) onto West African monsoon precipitation (GHCN), following removal of the linear variability associated with ENSO. Precipitation was averaged over the boxed region. Black contours delineate regions where the correlation coefficient is statistically significant at the 95% confidence level. Note the different color scales.
Fig. A5. Regression coefficients of modified versions of $\theta_{eb}$ (ERA-40) onto South Asian monsoon precipitation (GHCN). Color shading and black lines represent the regression coefficients (K[cm mo$^{-1}$]) and the 95% confidence level as in Fig. 2. Prior to regression analysis, $\theta_{eb}$ is calculated (a) with $T_b$ and subcloud pressure held constant, or (b) with $q_b$ and subcloud pressure held constant.