Observational Evaluation of a Convective Quasi-Equilibrium View of Monsoons

JI NIE, WILLIAM R. BOOS, AND ZHIMING KUANG

Department of Earth and Planetary Sciences, Harvard University, Cambridge, Massachusetts

(Manuscript received 5 November 2009, in final form 23 March 2010)

ABSTRACT

Idealized dynamical theories that employ a convective quasi-equilibrium (QE) treatment for the diabatic effects of moist convection have been used to explain the location, intensity, and intraseasonal evolution of monsoons. This paper examines whether observations of the earth's regional monsoons are consistent with the assumption of QE. It is shown here that in local summer climatologies based on reanalysis data, maxima of free-tropospheric temperature are, indeed, nearly collocated with maxima of subcloud equivalent potential temperature, θ_{eb}, in all monsoon regions except the North and South American monsoons. Free-tropospheric temperatures over North Africa also exhibit a strong remote influence from the South Asian monsoon. Consistent with idealized dynamical theories, peak precipitation falls slightly equatorward of the maxima in θ_{eb} and free-tropospheric temperature in regions where QE seems to hold.

Vertical structures of temperature and wind reveal two types of monsoon circulations. One is the deep, moist baroclinic circulation clearly seen in the South Asian monsoon. The other is of mixed type, with the deep moist circulation superimposed on a shallow dry circulation closely associated with boundary layer temperature gradients. While the existence of a shallow dry circulation has been documented extensively in the North African monsoon, here it is shown to also exist in Australia and southern Africa during the local summer. Analogous to moist QE theories for the deep circulation, the shallow circulation can be viewed in a dry QE framework in which shallow ascent occurs just equatorward of the peak boundary layer potential temperature, θ_b, providing a unified system where the poleward extents of deep and shallow circulations are bounded by maxima in θ_{eb} and θ_b, respectively.

1. Introduction

Monsoon circulations are important components of the general circulation of the atmosphere and have a large influence on climate and society in Asia, Africa, Australia, and the Americas (Ramage 1971; Webster 1987). These circulations are often viewed in a dry dynamical framework where precipitation over an off-equatorial land surface provides diabatic heating that drives thermally direct large-scale flow. Determining the distribution of precipitation and how it interacts with the large-scale flow, however, is one of the central unsolved problems of tropical meteorology. One traditional solution in monsoon dynamics is to specify precipitation as a function of low-level moisture convergence: low-level convergence is initially produced in early summer by the high temperatures achieved over a dry, off-equatorial land surface and is then maintained by the latent heating itself once monsoon precipitation has begun (e.g., Webster and Fasullo 2003).

An alternate treatment for precipitation used increasingly in recent decades is found in quasi-equilibrium (QE) theories of moist convection (Arakawa and Schubert 1974; Emanuel et al. 1994). These theories posit that moist, precipitating convection does not act as a heat source that drives large-scale flow, but instead simply relaxes the atmosphere toward a state of convective neutrality. In convecting regions, this results in covariation of the subcloud layer equivalent potential temperature, θ_{eb}, and the saturation value of this same quantity in the free troposphere, θ_e^*:

$$\delta\theta_{eb} \propto \delta\theta_e^*. \qquad (1)$$

Because the quantity θ_e^* is a saturation value, it depends only on temperature, while θ_{eb} is a function of both temperature and humidity. Quantities that influence the distribution of θ_{eb}, such as surface sensible and

Corresponding author address: Zhiming Kuang, Dept. of Earth and Planetary Sciences and School of Engineering and Applied Sciences, Harvard University, 20 Oxford St., Cambridge, MA 02138.
E-mail: kuang@fas.harvard.edu

DOI: 10.1175/2010JCLI3505.1

latent heat fluxes, can then be viewed as thermal forcings, and precipitation becomes an internal variable. Instead of equivalent potential temperatures, the QE relationship (1) is sometimes phrased in terms of moist entropy or moist static energy, both of which are closely related quantities approximately conserved in a moist atmosphere.

Numerous studies have used variations on this QE treatment of moist convection to conduct idealized theoretical investigations of monsoon dynamics (see reviews by Neelin 2007; Plumb 2007). For example, factors influencing the poleward extent and degree of zonal asymmetry in monsoons have been examined in the Quasi-Equilibrium Tropical Circulation Model (QTCM), a model containing two vertical modes in its dynamics and a simple quasi-equilibrium parameterization of moist convection (Chou et al. 2001; Chou and Neelin 2001, 2003). Dry theories of solsticial, axisymmetric Hadley circulations have been extended to a moist atmosphere by using relationships similar to (1) to reduce the vertical structure of the system. Emanuel (1995) showed that the critical temperature distribution derived by Plumb and Hou (1992) for the onset of nonlinear, angular momentum conserving meridional flow can be phrased in terms of a critical distribution of θ_{eb}, under the assumptions of QE.

Idealized theories of zonally symmetric Hadley circulations are of particular interest because they provide a clear picture of the relationship between the structure of meridional monsoon flow and the distributions of θ_{eb} and θ_e^* in a state of QE. In dry Hadley circulations subject to zonally symmetric thermal forcings that peak off the equator, the peak ascent zone is located just equatorward of the free-tropospheric temperature maximum (Lindzen and Hou 1988). This relationship seems to hold whether or not extratropical eddies play an important role in setting the strength of the overturning circulation, although the distance between the peak ascent zone and the thermal maximum does increase when the thermal maximum is moved as far poleward as 50° of latitude in eddy-permitting simulations (Walker and Schneider 2006). When the assumption of QE is added to these dynamical constraints on Hadley circulations, the free-tropospheric temperature maximum will be collocated with the maximum of θ_{eb} so that peak ascent and precipitation will occur just equatorward of the θ_{eb} maximum (or where $\partial\theta_{eb}/\partial y = 0$) (Emanuel 1995; Privé and Plumb 2007a).

An important point is that (1) is expected to hold only when deep moist convection is active. When a monsoon circulation exists, deep convection will be suppressed in regions of strong subsidence, and the free troposphere and subcloud layer will thermodynamically decouple in such regions. Because strong meridional overturning circulations generally consist of a broad region of subsidence and a narrow region of ascent and because the precipitating ascent zone is expected to be located near the free-tropospheric temperature maximum, the QE relationship (1) amounts to the condition that maxima of θ_{eb} and θ_e^* be collocated. In the rest of the domain, if subsidence is sufficiently strong, the two distributions may not covary, and the circulation itself is expected to weaken horizontal gradients of θ_e^*.

Although these relationships between the positions of maxima of precipitation, ascent, θ_{eb}, and θ_e^* have been shown to hold in simple models with parameterized convection and idealized boundary conditions (e.g., Privé and Plumb 2007b), the consistency with observations has not been comprehensively assessed. The validity of QE over tropical oceans in general was examined by Brown and Bretherton (1997), but that study largely excluded monsoon regions. Boos and Emanuel (2009) showed that the mean state of the South Asian summer monsoon is consistent with the idealized picture of QE monsoon dynamics discussed above, but such an assessment has not been performed for other monsoons. Performing this assessment is the main goal of this paper. To be clear, our focus is on one specific aspect of a QE view of monsoons, namely the alignment of the maxima in boundary-layer equivalent potential temperature and free-tropospheric saturation equivalent potential temperature. This is a necessary but not sufficient condition for the validity of models such as Emanuel (1995).

The QE framework discussed above assumes that moist, precipitating convection thermodynamically couples the subcloud layer with a deep layer of the troposphere. However, it is also well known that shallow, dry convection is very active over the Sahara and is associated with a shallow meridional circulation (Thorncroft and Blackburn 1999; Parker et al. 2005; Zhang et al. 2008). In this study, we show that qualitatively similar shallow circulations can be found in Australia and southern Africa during local summer. These shallow circulations can be viewed within a dry QE framework, where dry convection thermodynamically couples the temperature of the lower troposphere with that of near-surface air (Emanuel et al. 1994). The validity of this approach is also examined in this paper, and we show that several of the earth's regional monsoons consist of deep, moist circulations superimposed on dry, shallow circulations and that these circulations are respectively consistent with moist and dry QE frameworks.

The next section of this paper describes our methodology and data sources. Section 3 evaluates the validity of the QE hypothesis (1), in the dynamical context discussed above, for six major monsoon regions. In section 4, the vertical structure of these monsoons is analyzed, with particular emphasis on the shallow component of the circulation. The paper ends with a summary and discussion of open questions.

2. Data and methods

Four times daily estimates of wind, temperature, and humidity for 1979–2002 were obtained from the 40-yr European Centre for Medium-Range Weather Forecasts Re-Analysis (ERA-40) (Uppala et al. 2005), provided by the National Center for Atmospheric Research (NCAR). We used temperature, zonal wind, meridional wind, and vertical pressure velocity on pressure surfaces with a resolution of $2.5° \times 2.5°$. Temperature, humidity, and pressure on model level 57, a terrain-following level about 20 hPa above the ground surface, were used to estimate both the subcloud-layer equivalent potential temperature, θ_{eb}, and the subcloud-layer potential temperature, θ_b. This terrain-following data was obtained from the ERA-40 archive that was transformed by NCAR onto a 128×256 regular Gaussian grid. We recognize that reanalysis data may contain sizeable errors in regions where assimilated observations are sparse and this should be borne in mind when interpreting our results. However, while observations are likely sparse in the tropical oceanic subcloud layer, radiosonde observations are expected to constrain estimates of subcloud temperatures and humidities in many continental monsoon regions. Similarly, free-tropospheric temperatures are expected to be constrained by both radiosonde observations and, in more recent years, satellite soundings.

For estimates of precipitation, we used the areal averaged monthly accumulated rain data from the Tropical Rainfall Measuring Mission (TRMM, version 3B43v6) with $1° \times 1°$ resolution. The period of the data is about 9 years from January 1998 to September 2006.

All equivalent potential temperatures were calculated using the definitions in Emanuel (1994). Upper-tropospheric θ_e^* was calculated using a mass-weighted vertical average between 200 and 400 hPa, and all 6-hourly values within a month were averaged to obtain a monthly mean. When calculating θ_{eb}, special consideration was given to the possible influence of the diurnal cycle. In particular, instead of averaging daily mean values of θ_{eb}, daily maximum θ_{eb} was averaged over all days within each month. This was done because of the fact that both θ_{eb} and precipitation typically undergo high amplitude diurnal cycles over land in the tropics, and it seems likely that any thermodynamic coupling between the subcloud layer and the free troposphere will take place during episodes of precipitation, which will in turn presumably occur when θ_{eb} is high (although convective downdrafts may subsequently reduce θ_{eb}). Field campaign results support this hypothesis, showing diurnal fluctuations in θ_{eb} of over 10 K, with maxima of both precipitation and θ_{eb} occurring in the afternoon (Betts et al. 2002). Comparisons of climatologies of θ_{eb} produced using daily mean and daily maximum values show that, although the differences are not dramatic, the daily maximum θ_{eb} is more consistent with the θ_e^* distributions (see results in the next section).

3. Evaluation of horizontal structures

In this study, we focus on the mature stage of summer monsoons, when a strong meridional overturning circulation exists and sufficient precipitation should occur to thermodynamically couple the subcloud layer and upper troposphere. Monthly averaged θ_e^* and (daily maximum) θ_{eb} were calculated for July, August, January, and February. Figures 1 and 2 show θ_{eb}, θ_e^*, and precipitation in July (Northern Hemisphere summer monsoons) and January (Southern Hemisphere summer monsoons). Figures for August and February are similar with respect to the points discussed here and are therefore omitted.

We first discuss regional monsoons that exhibit a high degree of consistency with the QE hypothesis. For South Asia (Fig. 1a), maxima of θ_{eb} and θ_e^* are nearly collocated over northwestern India, as noted by Boos and Emanuel (2009) and Boos and Kuang (2010). The zonal offset between the two maxima is about 8° longitude, which seems quite small given the likely uncertainty in the data and the large horizontal scales over which the QE hypothesis is expected to hold. Although the maximum θ_{eb} is not positioned inside the 6 mm day^{-1} precipitation contour, it is inside the 3 mm day^{-1} precipitation contour (not shown). The Australian monsoon also exhibits maxima of θ_{eb} and θ_e^* that are very nearly collocated, with a departure of 6° in the zonal direction (Fig. 2a). In southern Africa there is no discernable offset between maxima of θ_{eb} and θ_e^* (Fig. 2b).

Southern Africa provides a good illustration of the importance of using the daily maximum θ_{eb} in monthly averages instead of the daily mean. The daily mean θ_{eb} was used to compute the climatology shown in Fig. 3; compared to Fig. 2b the inland maximum of θ_{eb} is considerably weaker, and the domain maximum is actually located over the ocean just west of Madagascar. Although the effect of using the daily maximum instead of the daily mean θ_{eb} is typically more minor in other regions, we have chosen to use daily maximum values for

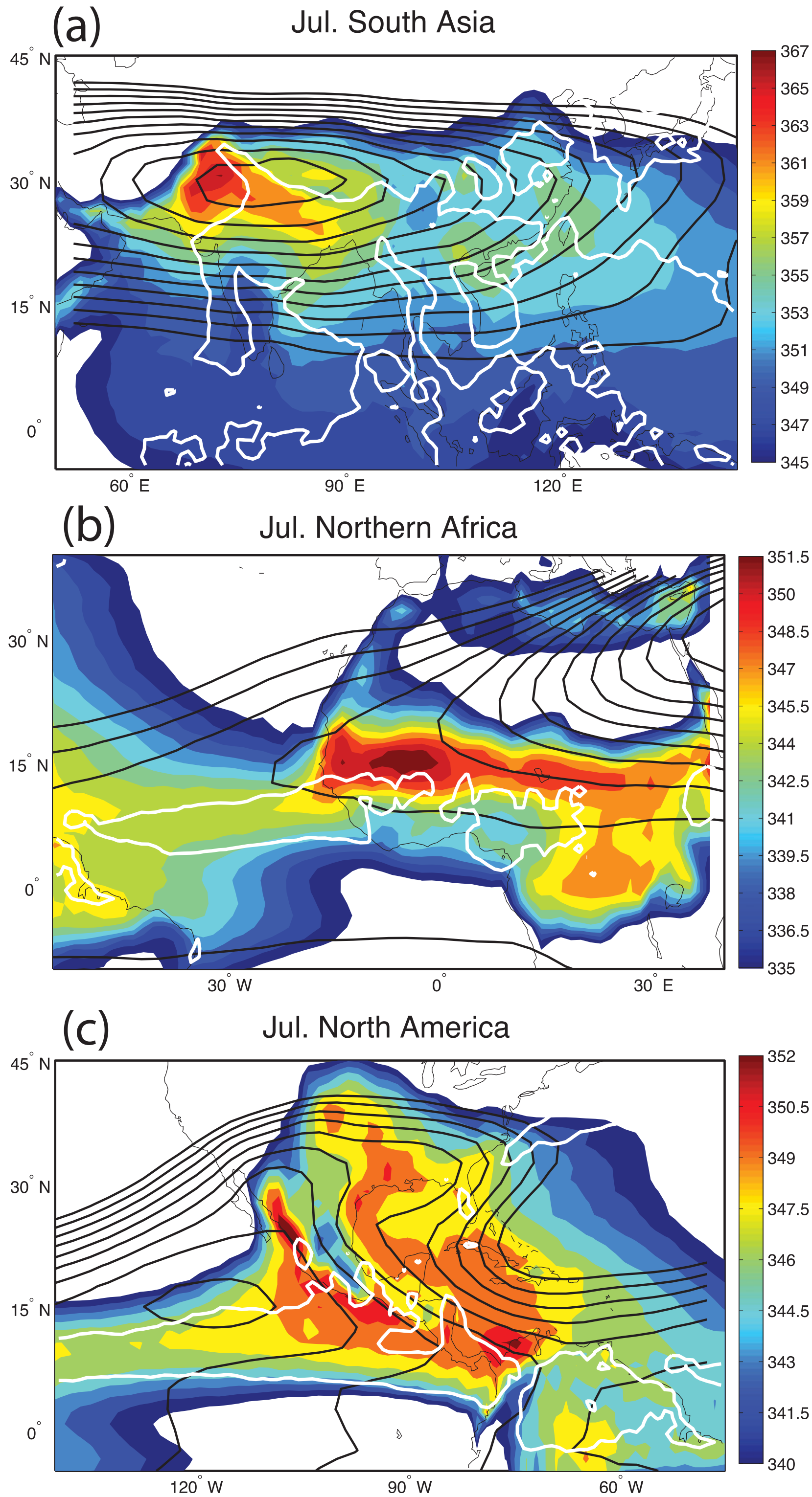

FIG. 1. Evaluation of QE for the (a) South Asia, (b) northern Africa, and (c) North America monsoons; subcloud θ_{eb} is shown. The black contour is the free tropospheric θ_e^* averaged from 200 to 400 hPa. The white contour indicates the region that has precipitation greater than 6 mm day^{-1}. The θ_e^* contours are started from (a) 345 K, (b) 340 K, and (c) 340 K and the respective interval is 1 K, 1 K, and 0.5 K.

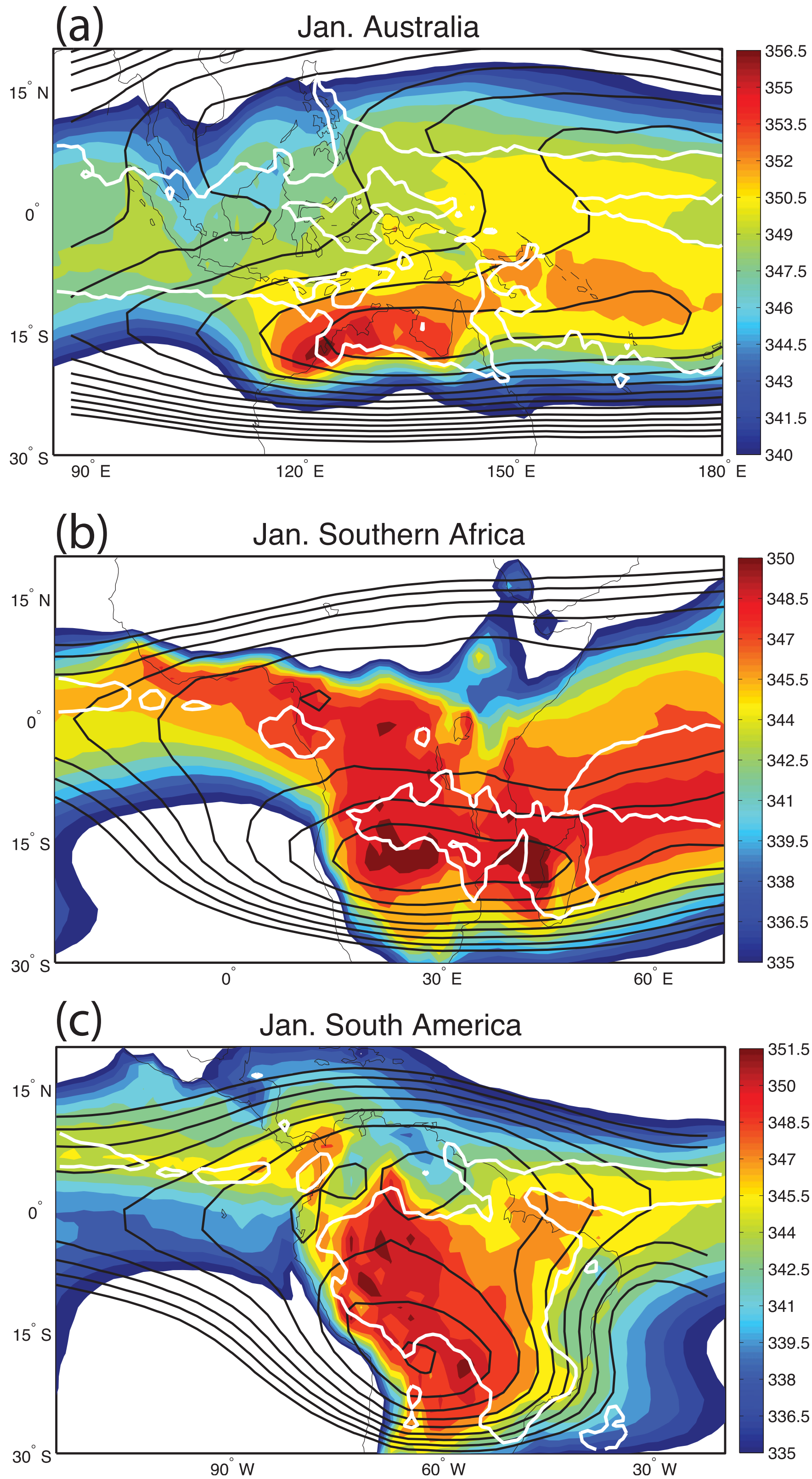

FIG. 2. As in Fig. 1 but for the (a) Australia, (b) southern Africa, and (c) South America monsoons. The θ_e^* contours start from 341 K at 0.5-K intervals.

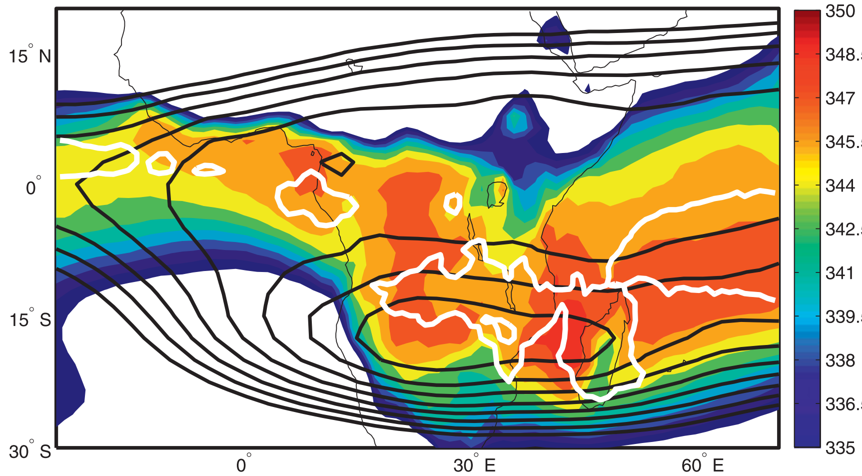

FIG. 3. As in Fig. 2b, but for daily mean θ_{eb}.

consistency with the theoretical view of the diurnal cycle of convection over land, discussed in the previous section.

Not all regional monsoons exhibit such a close correspondence between maxima of θ_{eb} and θ_e^*. The upper-tropospheric thermal structure over North Africa during local summer seems to be strongly influenced by the temperature maximum over South Asia (Fig. 1b). This is not surprising because a large region of subsidence is expected to exist over North Africa due to westward-traveling Rossby waves emitted by the South Asian monsoon (Rodwell and Hoskins 1996). However, the southwest–northeast tilted ridge of θ_e^* over North Africa suggests that there is a local influence from the belt of high θ_{eb} over the Sahel. Thus, the idealized picture of QE monsoon dynamics discussed in section 1 may hold in North Africa, with the added complication of a remote upper-tropospheric warming from the South Asian monsoon.

For North America, there are two regions of high θ_{eb}: over the Caribbean and over coastal western Mexico (Fig. 1c). The domain maximum θ_{eb} is near 20°N over western Mexico, more than 1000 km northeast of the θ_e^* peak, which is in turn located near the eastern Pacific intertropical convegence zone. Near 120°W in the eastern Pacific, a zonally elongated ridge in θ_e^* is closely aligned with a ridge in θ_{eb}, so it could be argued that QE is valid in a zonally averaged sense in this limited region. But, when considering the domain as a whole, there is clearly a large offset between the absolute maxima of θ_{eb} and θ_e^*. We also note that the Caribbean has rather high values of θ_{eb}, but relatively low values of precipitation and θ_e^*. Further analysis (not shown) suggests a possible explanation: the lower free troposphere is dry in this region, which could inhibit deep convection through entrainment and thus decouple the subcloud layer from the upper troposphere. The lower free troposphere might be dry in this region because of advection by the easterlies associated with the North Atlantic subtropical high. This suppression of precipitation by advection would be different from the ventilation mechanism proposed by Chou and Neelin (2003), which relies on advection to cool and dry the subcloud layer, rather than drying free-tropospheric air just above the subcloud layer.

In South America during local summer, the region of high θ_{eb} has a northwest–southeast tilt, somewhat parallel to topography of the continent's west coast. The maximum is extremely broad, with the highest values spreading from the equator to 20°S. The region of high θ_e^* also has a northwest–southeast tilt in the Southern Hemisphere and is not zonally elongated like it is in the other monsoon regions discussed above. The θ_e^* maximum is located near the poleward boundary of the highest θ_{eb} and just east of the Andes. The coincidence of both the θ_e^* maximum and sharp gradients of θ_{eb} with the Andes suggests that this topography may play an important role in setting the mean state of the South American monsoon, although cold SSTs in the Pacific might also have an effect. Some authors have concluded that the Bolivian high (which is roughly coincident with the upper-level θ_e^* maximum) is primarily a response to latent heating over the Amazon and that the Andes have little effect on this upper-tropospheric feature (Lenters and Cook 1997; Zhou and Lau 1998). However, these studies treat precipitation as an external heat source. If precipitation is, instead, viewed as an internal process in a QE framework, topography might play an important role in setting the characteristics of the Bolivian high by organizing the θ_{eb} distribution.

In summary, maxima of θ_{eb} and θ_e^* are closely collocated during local summer in South Asia, Australia, and South Africa. The QE picture may also be consistent with the North African monsoon if a remote upper-tropospheric forcing from the South Asian monsoon is taken into account, and in the East Pacific ITCZ. In South America, the peak θ_e^* is located at the poleward edge of a very broad θ_{eb} maximum, and it is unclear what sets the relative locations of these peaks. North America exhibits the most striking deviations from QE, with the maximum θ_{eb} positioned over coastal Mexico while the θ_e^* peak lies more than 1000 km to the southwest.

4. Evaluation of vertical structures

a. Thermodynamics

The previous section presented layer averages as a measure of the thermodynamic state of the free troposphere; this section examines the vertical structure of free-tropospheric θ_e^* in each of the monsoon regions discussed above. Regional zonal means of vertically resolved θ_e^* and zonal wind are plotted in Figs. 4 and 5 for each monsoon, together with θ_{eb}, θ_b, and precipitation averaged over the same longitudes. To more clearly distinguish local thermodynamic maxima, the distributions of θ_e^* are presented as anomalies relative to the mean value on each level. The maximum θ_e^* at each level is indicated by a black triangle.

In South Asia, θ_e^* maxima lie at approximately the same latitude as the θ_{eb} peak throughout the whole troposphere. Relative to the θ_{eb} maximum, there is a slight poleward bias of θ_e^* maxima on the upper slope of the Tibetan Plateau and at 200 and 300 hPa, but comparison with Fig. 1a shows that this may be due to the fact that the θ_{eb} distribution is less zonally symmetric than the θ_e^* distribution. In other words, the zonal mean operation biases the θ_{eb} peak slightly south of its position in latitude–longitude space. The vertically coherent free-tropospheric temperature structure is consistent with the monotonic change from westerlies to easterlies with height. This vertical structure is consistent with the circulation being dominated by a deep, moist Hadley circulation in this region, as discussed by numerous previous authors (e.g., Webster and Fasullo 2003). Consistent with the idealized meridional structure discussed in our introduction, the precipitation peak is located slightly equatorward of the θ_{eb} peak.

The vertical structure of other monsoons is quite different. In North Africa, θ_e^* in the lower troposphere (850–700 hPa) is maximum over the Sahara, about 8° north of the θ_{eb} peak. The low-level θ_e^* maximum is collocated with the θ_b maximum, as is expected in regions where dry

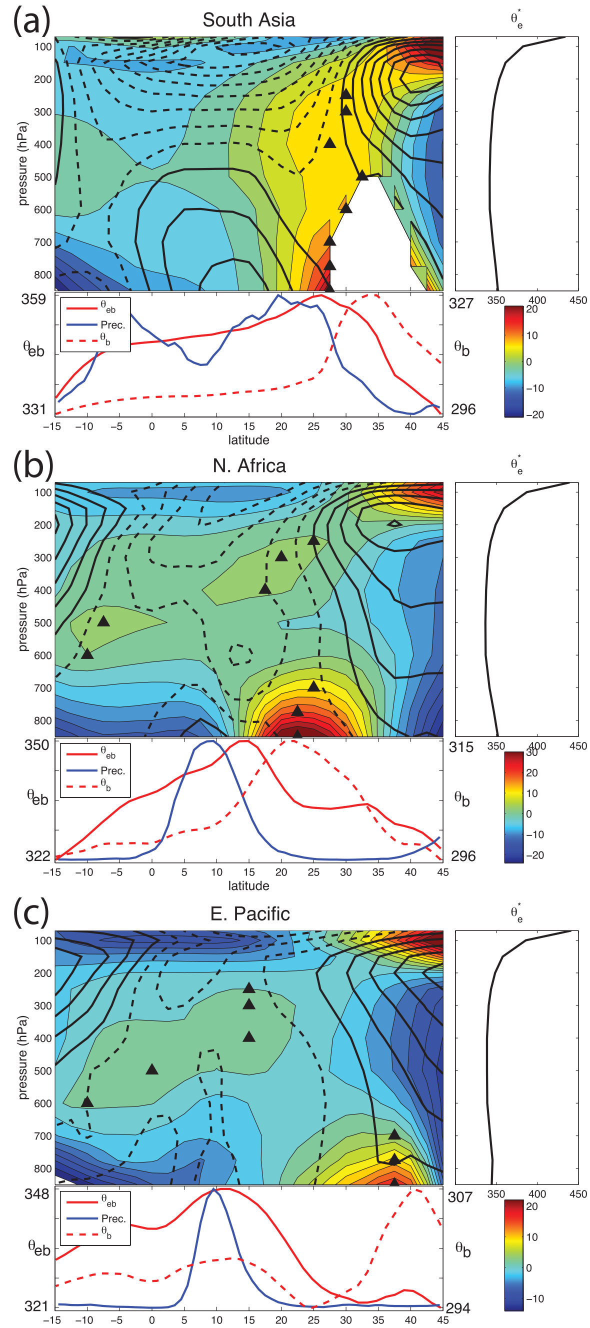

FIG. 4. July latitude–vertical cross sections of the (a) South Asia monsoon (averaged between 60° and 100°E), (b) Northern Africa monsoon (averaged between 20°W and 20°E), and (c) eastern Pacific ITCZ (averaged between 130° and 110°W). The upper-right panel is the domain-averaged θ_e^* on each level. For the upper-left panel, the color map is the θ_e^* anomaly with level mean (upper-right panel) removed. The color bar is in low-right corner. The black triangles indicate the θ_e^* maximum on each level; black contours indicate the zonal wind at contour interval 4 m s^{-1}; solid lines are westerlies, dashed lines easterlies, and the zero line is omitted. Lower panels depict subcloud θ_{eb}, subcloud θ_b, and precipitation. The precipitation peaks are (a) 8.77, (b) 7.70, and (c) 13.78 mm day^{-1}.

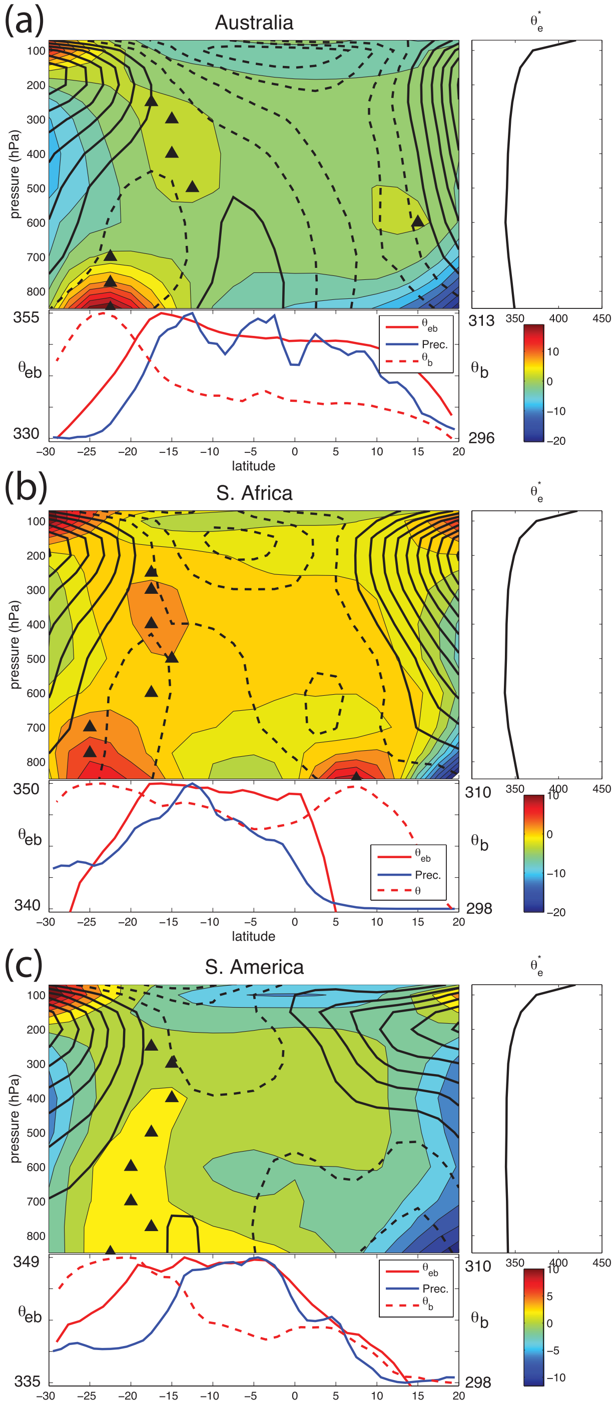

FIG. 5. As in Fig. 4 but in January for the (a) Australia monsoon (averaged between 120° and 140°E), (b) southern Africa monsoon (averaged between 15° and 35°E), and (c) South America monsoon (averaged between 80° and 50°W). Precipitation peaks are (a) 8.56, (b) 7.16, and (c) 8.46 mm day^{-1}.

convection vertically homogenizes potential temperature in a boundary layer confined to the lower troposphere. This distribution of θ_e^* is associated with near-equatorial surface westerlies and a lower-tropospheric easterly jet centered near 600 hPa (see also Thorncroft and Blackburn 1999). The wind becomes more westerly in the 400~500-hPa layer; then the vertical shear changes sign yet again to produce stronger easterlies in the upper troposphere, consistent with the reversals of the meridional temperature gradient in this region. Thorncroft and Hoskins (1994) discussed how the lower-tropospheric easterly jet is unstable and generates waves that converge westerly angular momentum into the jet. It is the equatorward return flow of the shallow circulation that maintains the easterly jet by advecting easterly planetary momentum into its core (Thorncroft and Blackburn 1999).

In the middle troposphere (600~500 hPa) in these West African longitudes, the θ_e^* maximum lies in the winter hemisphere. This is consistent with a "cold top" lying above a warm layer; cold temperatures above the shallow warm layer are generated by adiabatic cooling due to lifting as the warm layer expands (Holloway and Neelin 2007). In the upper troposphere, the θ_e^* distribution seems to be influenced both by subcloud layer properties and remote warming from the South Asian monsoon. The θ_e^* maximum at 400 hPa lies at nearly the same latitude as the θ_{eb} maximum, but the peak θ_e^* then shifts poleward with height. Association of this poleward shift with the Asian monsoon is more clearly inferred from Fig. 1b. Note that the extratropical maximum in θ_e^* near 100 hPa is associated with stratospheric temperatures and the decreasing height of the tropopause with latitude and is, thus, not directly relevant to the theoretical frameworks examined here.

The pattern of the maxima in θ_e^* lying in the winter hemisphere near 600 hPa is clearly seen in the eastern Pacific, in a zonal mean over the region where QE seemed to approximately hold (Fig. 4c). Note that there are two local maxima in θ_b in this region, one near the ITCZ at 12°N and another over the Mojave Desert farther poleward. The shallow circulation associated with the 600-hPa temperature maximum in the winter hemisphere is likely associated with the ITCZ and has been discussed by previous authors (e.g., Zhang et al. 2004). In this sense, this figure better describes the eastern Pacific ITCZ than the North America monsoon. Indeed, many issues discussed here are similar to those discussed in the context of finding the distribution of precipitation given the SST—in particular, the relative roles of dynamic considerations, such as boundary layer pressure (set by θ_b), and thermodynamic considerations, such as boundary layer moist static energy distributions (Lindzen and Nigam 1987; Sobel and Neelin

2006; Sobel 2007; Back and Bretherton 2009). This work simply extends our examination of these issues to monsoon regions.

Although numerous studies have documented the existence of a shallow dry circulation in North Africa and the eastern Pacific (see Zhang et al. 2008 for a review), there are few reports of similar shallow circulations existing during mean summer in other monsoons. Shallow circulations have been documented prior to the monsoon onset in Asia and Australia, but are typically thought to give way to deep flow once the summer monsoon begins (Webster 1987). However, a winter hemisphere maximum in θ_e^* at 600 hPa is seen in a zonal mean over Australian longitudes during January (Fig. 5a). At lower altitudes the θ_e^* maxima are closely collocated with the θ_b peak, and a weak easterly maximum exists at low levels near 20°S. In the upper troposphere, θ_e^* maxima are positioned closer to the equator, nearly collocated with the θ_{eb} peak. The majority of precipitation lies equatorward of the θ_{eb} peak, although the precipitation maximum is much broader than suggested by the idealized picture of moist Hadley circulations described in the introduction. This seems to be true for the South Asian and South American monsoons as well. Overall, these patterns suggest that the Australian monsoon exhibits a similar structure to the North African monsoon with a shallow, dry circulation superimposed on the deep, moist circulation.

In southern Africa (Fig. 5b), the low-level thermodynamic structure is more complicated. In the Southern Hemisphere, the Kalahari Desert lies poleward of the peak precipitation and is associated with maxima in θ_b and low-level θ_e^* near 25°S. In the Northern Hemisphere, the North African savanna belt (5°∼10°N, 15°W∼35°E) has high boundary layer temperatures even in winter. These local maxima in low-level temperature are each associated with weak easterly jets near 600 hPa, suggesting that a shallow, dry circulation might exist on each side of the equator. In the upper troposphere, maxima in θ_e^* are nearly collocated with the θ_{eb} maximum near 15°S, and the peak precipitation falls slightly equatorward of that latitude.

Limited zonal means are more difficult to interpret for South America (Fig. 5c) because of the northwest–southeast tilt of the thermodynamic maxima. Nevertheless, a zonal mean from 50° to 80°W shows that the upper-tropospheric θ_e^* maximum is located at the poleward edge of a broad region of high θ_{eb}, consistent with the discussion in the previous section. When calculating Fig. 5c, values of θ_{eb} and θ over the ocean are excluded because of the tilted and strong gradient (Fig. 2c). The free-tropospheric θ_e^* field is fairly coherent in the vertical; although there is some poleward bias of the θ_e^* maxima at lower levels compared to upper levels, meridional gradients are fairly weak in general in this region.

Based on the observational analyses presented above, circulations in the six regions that we examined can be separated into two types. One has only a deep, moist circulation with an ascent branch and a precipitation maximum located just equatorward of the θ_{eb} peak (Fig. 6a). This seems to be a decent description of the South Asian monsoon. In other regions a shallow, dry circulation is superimposed on the deep, moist circulation with the ascent branch of the shallow circulation located near the θ_b maximum, which in turn is typically located poleward of the θ_{eb} maximum (Fig. 6b). The shallow poleward temperature gradients are associated with a low-level easterly jet and an equatorward temperature gradient in middle levels. This mixed circulation seems to describe important elements of the summer state in North Africa, the eastern Pacific, Australia, and southern Africa.

b. Winds

Figures 7 and 8 present the reanalyzed meridional circulation for the same limited zonal means, in order to confirm that the structures discussed above can be seen directly in an estimate of the flow field. Meridional and vertical winds are presented as vectors because mass is not conserved in a limited zonal mean, o a mass streamfunction cannot be defined. Upward and downward motion is plotted using different colors to highlight the rising branch of any shallow circulations. The latitudes of maxima of θ_{eb} and θ_b are marked with vertical solid and dashed lines, respectively, to aid comparison with the QE framework. Data at pressures greater than surface pressure are masked. Although ERA-40 data is presented here, this analysis was repeated with the National Centers for Environmental Prediction (NCEP)–NCAR Global Reanalysis 1 (Kalnay et al. 1996). Minor differences were found between these two datasets, consistent with results reported by Zhang et al. (2008), and the results discussed below were overall very similar.

In South Asia (Fig. 7a) there is only a deep circulation, with the θ_{eb} maximum positioned approximately at the poleward boundary of the circulation's rising branch. Some ascent does occur over the elevated topography near 30°N, but this has a small contribution to the total overturning mass flux.

In North Africa (Fig. 7b), Australia (Fig. 8a), and southern Africa (Fig. 8b), shallow ascent can clearly be seen poleward of the maximum deep ascent. The meridional structure is largely consistent with that outlined in the introduction, with peak deep ascent occurring just

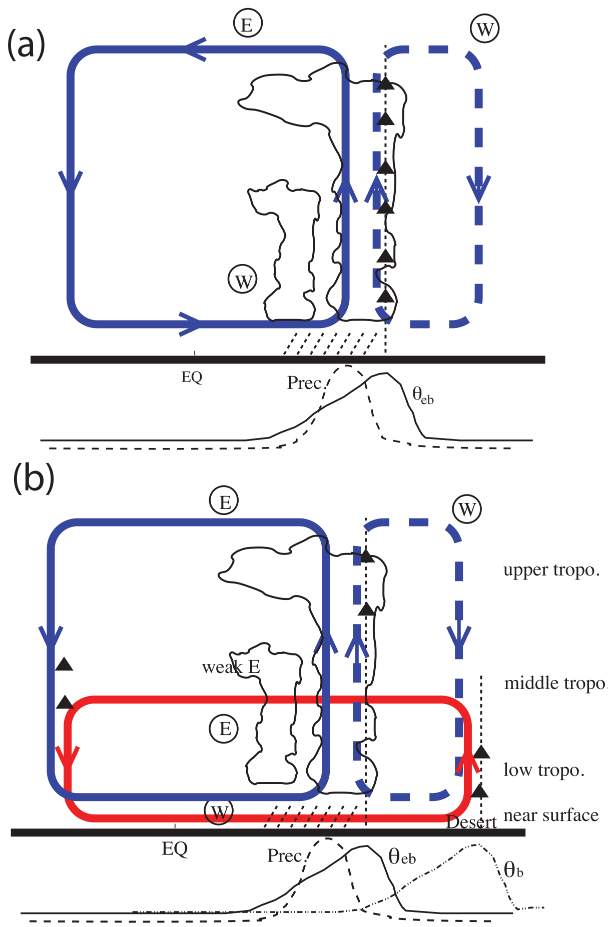

FIG. 6. Diagrams of zonal mean monsoon circulations of (a) deep and moist Hadley cell type and (b) mixed type for a vertical–meridional cross section. The solid and dashed boldface lines indicate the circulation for the Northern Hemisphere. Jet cores are shown as "E" or "W" in the black circle; black triangles indicate θ_e^* peaks on each pressure level; θ_{eb}, θ_b, and precipitation are plotted in the same manner as in Figs. 4 and 5.

equatorward of the θ_{eb} maximum and peak shallow ascent occurring just equatorward of the θ_b maximum. As inferred above, shallow ascent occurs on both sides of the equator during local summer in southern Africa. In North Africa the shallow circulation actually crosses the equator in the midtroposphere (near 600 hPa), while in Australia and South Africa it feeds into the deep ascent, and cross-equatorial flow occurs in the upper troposphere. This may be an important difference between the shallow circulations in these regions, so the schematic in Fig. 6b may need to be modified to show the low-level cell confined to the summer hemisphere for Australia and southern Africa.

For the East Pacific (Fig. 7c), any shallow ascent seems to occur at the same latitude as the deep ascent. This is consistent with maxima of θ_{eb} and θ_b being

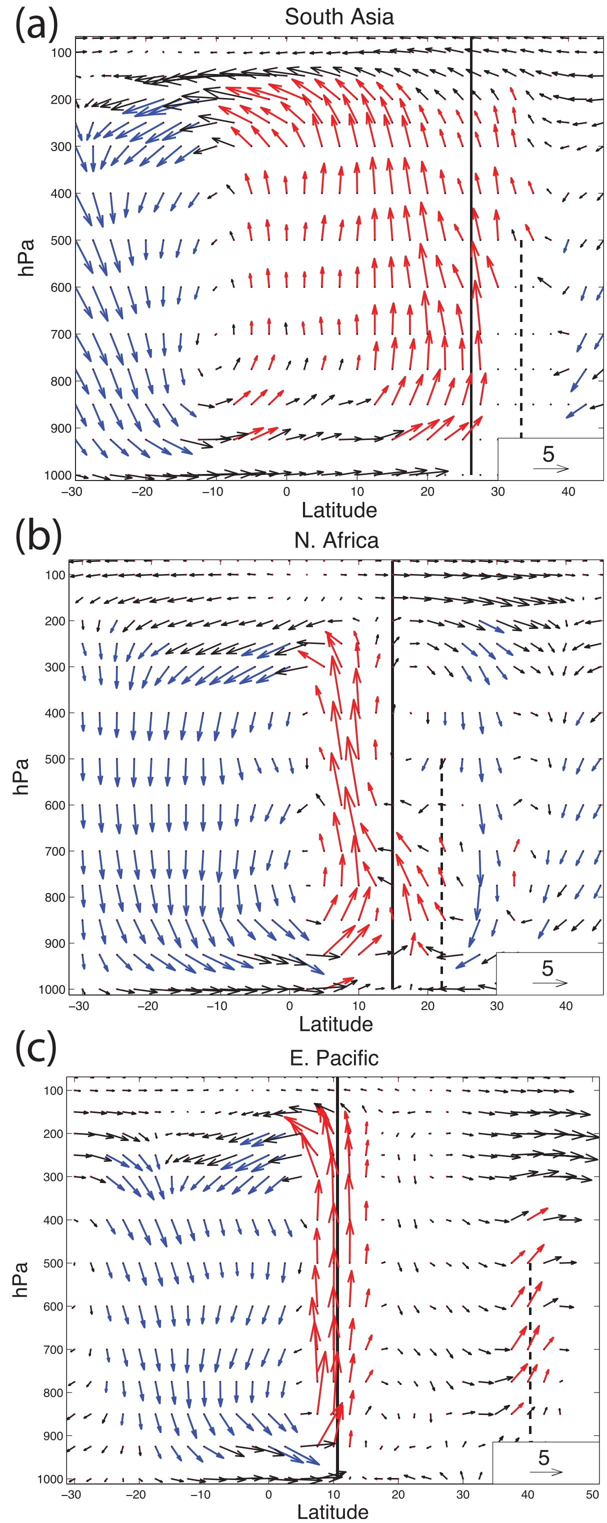

FIG. 7. July latitude–vertical cross section of the wind vectors of the (a) South Asia monsoon, (b) northern Africa monsoon, and (c) eastern Pacific ITCZ. Vectors with upward (downward) motion stronger than -0.02 Pa s^{-1} (0.02 Pa s^{-1}) are red (blue); other vectors are black. The vertical black solid (dotted) line indicates the latitude of the θ_{eb} (θ_b) peak. The vertical velocity is amplified 80 times for better illustration.

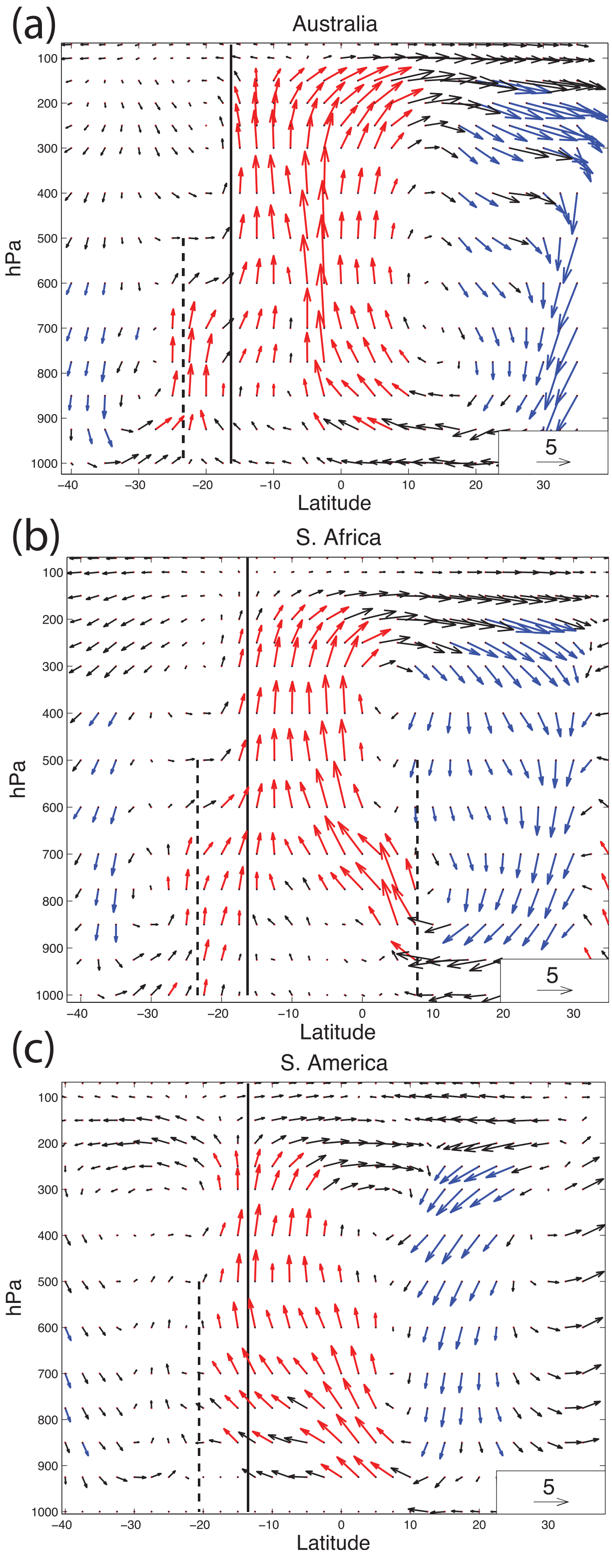

FIG. 8. As in Fig. 7 but for January during the (a) Australia, (b) southern Africa, and (c) South America monsoons.

collocated over the warmest SST in this region, as was shown at the bottom of Fig. 4c. Shallow ascent also occurs near the second θ_b maximum at 40°N, but the flow in this region is poleward rather than equatorward. The wind composite of South America (Fig. 8c) shows a single deep rising branch. However, the θ_{eb} maximum is very broad in this region, so it is difficult to assess the alignment of the poleward extension of the rising branch and the latitude of the θ_{eb} maximum. The results here are consistent with the previous analyses shown in Figs. 2c and 5c.

5. Summary and discussion

A new framework of monsoon dynamics has emerged in recent years based on idealized theories using convective quasi-equilibrium: the maximum equivalent potential temperature in the subcloud layer is collocated with the maximum free-tropospheric temperature, and the main precipitating ascent zone of the cross-equatorial monsoon flow is located just equatorward of these maxima (e.g., Emanuel 1995; Privé and Plumb 2007a). Our main goal in this paper was to evaluate the consistency of this picture with observed monsoons, a task that had not been performed despite the fact that numerous studies have used QE treatments of moist convection in idealized dynamical models (see reviews by Plumb 2007; Neelin 2007).

In South Asia, Australia, and southern Africa during local summer, maxima of upper-tropospheric θ_e^* were quite closely collocated with θ_{eb}. In North Africa, upper-tropospheric θ_e^* seemed to be influenced by the local θ_{eb} distribution as well as by a remote warming from the South Asian monsoon. While the QE relationship (1) seems to hold in a limited zonal mean over the East Pacific ITCZ, it does not seem to hold in the North American monsoon. South America also exhibited some departure from this QE relationship, as the peak upper-level θ_e^* was highly localized at the poleward edge of a very broad maximum high θ_{eb}.

Vertical structures of both temperature and wind fields suggest that there are two broad categories of meridional monsoon circulations. One is the classic deep, moist circulation, exemplified by the South Asian monsoon. The other is of mixed type, with a dry, shallow circulation superimposed on the deep, moist circulation. The two circulations might be thought of as existing in states of dry and moist QE, respectively, with deep moist convection coupling upper-tropospheric θ_e^* to θ_{eb} in the deep circulation, and shallow dry convection coupling lower-tropospheric θ_e^* to θ_b in the shallow circulation. The rising branches of the deep and shallow circulations are thus expected to be bounded on their poleward side by maxima of θ_{eb} and θ_b, respectively. While previous work has documented this sort of mixed circulation in the North African monsoon and the East Pacific ITCZ (e.g., Zhang et al. 2008), our contribution here is to show that qualitatively similar circulations seem to exist in Australia and southern Africa. Important regional differences do exist, however, with the shallow circulations in Australia and southern Africa producing little cross-equatorial flow in the midtroposphere.

The convective QE framework has the potential to simplify our understanding of monsoon dynamics, but it leaves open the question of what sets the distributions of θ_{eb} and θ_b. Although theories such as subcloud-layer quasi-equilibrium (Raymond 1995) seek to provide additional constraints on these distributions, the ways in which surface enthalpy fluxes, radiation, horizontal advection, and convective downdrafts interact to set the thermal properties of the subcloud layer merit further examination, especially within the context of monsoons.

It is of interest to consider the role of the topography in the South Asian and South American monsoons, and its possible relevance to the existence of a shallow circulation. In the absence of elevated topography, the precipitating ascent zone in South Asia has been shown to shift equatorward and decrease in intensity (Hahn and Manabe 1975; Prell and Kutzbach 1992). This change is accompanied by the formation of a broad desert in the current location of the Tibetan Plateau. Such a desert might produce a shallow dry circulation and thus change the South Asian monsoon from a pure deep and moist circulation to a circulation of mixed type. In other words, the mixed circulation may be the more general type of monsoon flow, with the pure deep circulation occurring only when the dynamics are modified by topography or other factors.

The coexistence of shallow and deep circulations also leads to an interesting question: do these circulations interact with each other? Although little work has been done to examine this question, some relevant research shows hints of an interaction. Both models and observations show that precipitation over the Sahel is positively correlated with the strength of the Sahara low (Biasutti et al. 2009), which suggests that such interaction may occur. Transitions between shallow and deep flow also take place within each season, with considerable complexity in the seasonal onset of the North African monsoon (Gu and Adler 2004). Understanding the mechanisms by which shallow and deep circulations interact thus seems a useful direction for future research.

Acknowledgments. We thank Kerry Emanuel very much for valuable comments and careful corrections. This research is partially supported by NSF Grant ATM-0754332. William R. Boos was supported by the Reginald A. Daly Postdoctoral Fellowship in the Department of Earth and Planetary Sciences at Harvard University, and the John and Elaine French Environmental Fellowship at the Harvard University Center for the Environment.

REFERENCES

Arakawa, A., and W. Schubert, 1974: Interaction of a cumulus cloud ensemble with the large-scale environment, Part I. *J. Atmos. Sci.,* **31,** 674–701.

Back, L. E., and C. S. Bretherton, 2009: A simple model of climatological rainfall and vertical motion patterns over the tropical oceans. *J. Climate,* **22,** 6477–6497.

Betts, A. K., J. D. Fuentes, M. Garstang, and J. H. Ball, 2002: Surface diurnal cycle and boundary layer structure over Rondônia during the rainy season. *J. Geophys. Res.,* **107,** 8065, doi:10.1029/2001JD000.

Biasutti, M., A. H. Sobel, and S. J. Camargo, 2009: The role of the Sahara low in summertime Sahel rainfall variability and change in the CMIP3 models. *J. Climate,* **22,** 5755–5771.

Boos, W. R., and K. A. Emanuel, 2009: Annual intensification of the Somali jet in a quasi-equilibrium framework: Observational composites. *Quart. J. Roy. Meteor. Soc.,* **135,** 319–335.

——, and Z. Kuang, 2010: Dominant control of the South Asian monsoon by orographic insulation versus plateau heating. *Nature,* **463,** 218–222.

Brown, R., and C. Bretherton, 1997: A test of the strict quasi-equilibrium theory on long time and space scales. *J. Atmos. Sci.,* **54,** 624–638.

Chou, C., and J. Neelin, 2001: Mechanisms limiting the southward extent of the South American summer monsoon. *Geophys. Res. Lett.,* **28,** 2433–2436.

——, and ——, 2003: Mechanisms limiting the northward extent of the northern summer monsoons over North America, Asia, and Africa. *J. Climate,* **16,** 406–425.

——, ——, and H. Su, 2001: Ocean–atmosphere–land feedbacks in an idealized monsoon. *Quart. J. Roy. Meteor. Soc.,* **127,** 1869–1892.

Emanuel, K. A., 1994: *Atmospheric Convection*. Oxford University Press, 580 pp.

——, 1995: On thermally direct circulations in moist atmospheres. *J. Atmos. Sci.,* **52,** 1529–1536.

——, J. D. Neelin, and C. S. Bretherton, 1994: On large-scale circulations in convecting atmospheres. *Quart. J. Roy. Meteor. Soc.,* **120,** 1111–1143.

Gu, G., and R. F. Adler, 2004: Seasonal evolution and variability associated with the West African monsoon system. *J. Climate,* **17,** 3364–3377.

Hahn, D. G., and S. Manabe, 1975: The role of mountains in the South Asian monsoon circulation. *J. Atmos. Sci.,* **32,** 1515–1541.

Holloway, C., and J. Neelin, 2007: The convective cold top and quasi equilibrium. *J. Atmos. Sci.,* **64,** 1467–1487.

Kalnay, E., and Coauthors, 1996: The NCEP/NCAR 40-Year Reanalysis Project. *Bull. Amer. Meteor. Soc.,* **77,** 437–471.

Lenters, J. D., and K. H. Cook, 1997: On the origin of the Bolivian high and related circulation features of the South American climate. *J. Atmos. Sci.,* **54,** 656–677.

Lindzen, R. S., and S. Nigam, 1987: On the role of sea surface temperature gradients in forcing low-level winds and convergence in the tropics. *J. Atmos. Sci.,* **44,** 2418–2436.

——, and A. Y. Hou, 1988: Hadley circulation for zonally averaged heating centered off the equator. *J. Atmos. Sci.,* **45,** 2416–2427.

Neelin, J., 2007: Moist dynamics of tropical convection zones in monsoons, teleconnections and global warming. *The Global Circulation of the Atmosphere,* T. Schneider and A. H. Sobel, Eds., Princeton University Press, 267–301.

Parker, D. J., C. D. Thorncroft, R. R. Burton, and A. Diongue-Nian, 2005: Analysis of the African easterly jet, using aircraft observations from the JET2000 experiment. *Quart. J. Roy. Meteor. Soc.,* **131,** 1461–1482.

Plumb, R. A., 2007: Dynamical constraints on monsoon circulations. *The Global Circulation of the Atmosphere,* T. Schneider and A. H. Sobel, Eds., Princeton University Press, 252–266.

——, and A. Y. Hou, 1992: The response of a zonally symmetric atmosphere to subtropoical thermal forcing: Threshold behavior. *J. Atmos. Sci.,* **49,** 1790–1799.

Prell, W., and J. Kutzbach, 1992: Sensitivity of the Indian monsoon to forcing parameters and implications for its evolution. *Nature,* **360,** 647–652.

Privé, N. C., and R. A. Plumb, 2007a: Monsoon dynamics with interactive forcing. Part I: Axisymmetric studies. *J. Atmos. Sci.,* **64,** 1417–1430.

——, and ——, 2007b: Monsoon dynamics with interactive forcing. Part II: Impact of eddies and asymmetric geometries. *J. Atmos. Sci.,* **64,** 1431–1442.

Ramage, C. S., 1971: *Monsoon Meteorology*. Academic Press, 296 pp.

Raymond, D. J., 1995: Regulation of moist convection over the West Pacific warm pool. *J. Atmos. Sci.,* **52,** 3945–3959.

Rodwell, M., and B. Hoskins, 1996: Monsoons and the dynamics of deserts. *Quart. J. Roy. Meteor. Soc.,* **122,** 1385–1404.

Sobel, A. H., 2007: Simple models of ensemble-averaged precipitation and surface wind, given the SST. *The Global Circulation of the Atmosphere,* T. Schneider and A. H. Sobel, Eds., Princeton University Press, 219–251.

——, and J. D. Neelin, 2006: The boundary layer contribution to intertropical convergence zones in the quasi-equilibrium tropical circulation model framework. *Theor. Comput. Fluid Dyn.,* **20,** 323–350.

Thorncroft, C. D., and B. J. Hoskins, 1994: An idealized study of African easterly waves. Part 1: A linear view. *Quart. J. Roy. Meteor. Soc.,* **120,** 953–982.

——, and M. Blackburn, 1999: Maintenance of the African easterly jet. *Quart. J. Roy. Meteor. Soc.,* **125,** 763–786.

Uppala, S., and Coauthors, 2005: The ERA-40 re-analysis. *Quart. J. Roy. Meteor. Soc.,* **131,** 2961–3012.

Walker, C. C., and T. Schneider, 2006: Eddy influences on Hadley circulations: Simulations with an idealized GCM. *J. Atmos. Sci.,* **63,** 3333–3350.

Webster, P. J., 1987: The elementary monsoon. *Monsoons,* J. S. Fein and P. L. Stephens, Eds., John Wiley, 3–32.

——, and J. Fasullo, 2003: Monsoon: Dynamical theory. *Encyclopedia of Atmospheric Sciences,* J. R. Holton, Ed., Academic Press, 1370–1385.

Zhang, C., M. McGauley, and N. Bond, 2004: Shallow meridional circulation in the tropical eastern Pacific. *J. Climate,* **17,** 133–139.

——, D. S. Nolan, C. D. Thorncroft, and H. Nguyen, 2008: Shallow meridional circulations in the tropical atmosphere. *J. Climate,* **21,** 3453–3470.

Zhou, J., and K. Lau, 1998: Does a monsoon climate exist over South America? *J. Climate,* **11,** 1020–1040.