### Dynamics of the Shallow Meridional Circulation around Intertropical Convergence Zones

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#### ABSTRACT

The generally accepted view of the meridional circulation in the tropical east Pacific is that of a single deep overturning cell driven by deep convective heating in the intertropical convergence zone (ITCZ), similar to the zonal mean Hadley circulation. However, recent observations of the atmosphere from the tropical eastern Pacific have called this view into question. In several independent datasets, significant meridional return flows out of the ITCZ region were observed, not only at high altitudes, but also at low altitudes, just above the atmospheric boundary layer. This paper presents a theory and idealized simulations to understand the causes and dynamics of this shallow meridional circulation (SMC).

Fundamentally, the SMC can be seen as a large-scale sea-breeze circulation driven by sea surface temperature gradients when deep convection is absent in the ITCZ region. A simple model of this circulation is presented. Using observed values, the sea-breeze model shows that the pressure gradient above the boundary can indeed reverse, leading to the pressure force that drives the shallow return flow out of the ITCZ.

The Weather Research and Forecast Model (WRF) is used to simulate an idealized Hadley circulation driven by moist convection in a tropical channel. The SMC is reproduced, with reasonable similarity to the circulation observed in the east Pacific. The simulations confirm that the SMC is driven by a reversal of the pressure gradient above the boundary layer, and that the return flow is strongest when deep convection is absent in the ITCZ, and weakest when deep convection is active. The model also shows that moisture transport out of the ITCZ region is far greater in the low-level shallow return flow than in the high-altitude return flow associated with the deep overturning, and that a budget for water transport in and out of the ITCZ region is grossly incomplete without it. Much of the moisture carried in the shallow return flow is recycled into the boundary layer, but does not appear to contribute to enhanced cloudiness in the sub-tropical stratocumulus poleward of the ITCZ.

#### 1. Introduction

Textbooks describe the meridional circulation pattern in the east Pacific as dominated by a deep overturning circulation similar to the zonal-mean Hadley circulation (e.g., Peixoto and Oort 1992; Grotjahn 1993). The ascending branch of this deep circulation resides in the intertropical convergence zone (ITCZ),

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which in the east Pacific remains north of the equator for the entire year, while its descending branch remains south of the equator. These two branches of vertical motion are connected by the trade wind southerly flow in the atmospheric boundary layer and the northerly return flow in the upper troposphere. The vertical and upper branches of this deep meridional circulation, which will hereafter be referred to simply as the deep circulation, is sketched as the dashed lines in Fig. 1.

Recent scrutiny of in situ observations, however, has revealed that there exists a shallow meridional circulation in the tropical east Pacific (Zhang et al. 2004). This circulation shares the same boundary layer southerly flow with the deep circulation and its ascending branch is also rooted in the ITCZ. However, the northerly re-

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FIG. 1. An idealized view of the deep and shallow overturnings in the east Pacific, taken from Zhang et al. (2004).

turn flow of this shallow circulation is located immediately above the boundary layer. The vertical extent of the return flow ranges from 2–4 km. This shallow meridional circulation (SMC), is sketched as the solid lines in Fig. 1; the flow out of the ITCZ just above the boundary layer will be referred to as the shallow return flow (SRF).

The SMC has been observed in four independent, in situ datasets: First Global Atmospheric Research Program (GARP) Global Experiment (FGGE) dropsondes (Yin and Albrecht 2000), soundings launched from ships tending the Tropical Atmosphere–Ocean (TAO) mooring array (Bond 1992), East Pacific Investigation of Climate Processes in the Coupled Ocean– Atmosphere System 2001 (EPIC2001; see Raymond et al. 2004) dropsondes, and wind profiler data from Christmas Island (2°N, 157.4°W) and San Cristóbal, Galápagos (0.9°S, 89.6°W; Gage et al. 1994).

Murakami et al. (1992) speculated the possibility of an SMC from examining the depth of convection in the Tropics. The SMC has also been indicated by the divergent wind component of global model reanalysis products (Tomas and Webster 1997; Trenberth et al. 2000). Its existence was confirmed only after examination of the in situ observations cited above. In fact, further analysis has shown the SMC is stronger in the observations than in the global reanalyses (Zhang et al. 2006, manuscript submitted to *J. Climate*). The SRF in the east Pacific undergoes a distinct seasonal cycle. It is the strongest in boreal fall to early winter, and the weakest in boreal spring (Zhang et al. 2004).

The existence of the SMC is consistent with the notion that convection and dynamics in the Tropics has two modes: one associated with shallow convection, and the other with deep convection. In an empirical orthogonal function analysis of the vertical structure of the divergent wind in global model reanalysis data, Trenberth et al. (2000) found that the first leading mode in the east Pacific represents the deep circulation and the second leading mode represents the SMC. High-resolution numerical simulations have shown that in highly idealized environments convection will spontaneously organize into clusters of deep convection and areas of shallow or no convection (Tompkins and Craig 1998; Tompkins 2001; Bretherton et al. 2005).

A feature quite similar to the observed SMC was previously seen in highly idealized simulations by Schneider and Lindzen (1977), Schneider (1977), and to some extent by Held and Hou (1980). All three studies used zonally symmetric models of incompressible flow, with fixed stratification, and simple radiative parameterizations based on Newtonian relaxation to prescribed atmospheric temperature profiles. The important distinction is that the Schneider and Lindzen simulations used a meridionally varying temperature as a lower boundary condition, while Held and Hou used a no-flux lower boundary condition. While the simulations of Held and Hou only hinted at a low-level circulation, the simulations of Schneider and Lindzen produced a distinct low-level overturning, restricted below 800 hPa, quite similar to the SMC discussed here. Schneider and Lindzen (1977) postulated that this circulation was driven by the SST gradient, and it is this hypothesis upon which we will expand.

Shallow meridional circulations have also appeared in more recent numerical simulations of convection in nonrotating environments with zonal temperature gradients. Grabowski et al. (2000) used a two-dimensional, cloud-resolving model with 1.8-km horizontal resolution to simulate a Walker-type circulation driven by an SST gradient in the zonal direction. The simulations produced two distinct circulations in the vertical, with a low-level return flow emanating out of the convecting region. Larson and Hartmann (2003) also simulated a Walker-type circulation, but used a radically different model, with a three-dimensional periodic channel, grid spacing of 120 km, and cumulus parameterization. When the horizontal SST gradient was sufficiently large (8 K over 9600 km), a shallow (zonal) circulation also appeared beneath the deep circulation. While it should be noted that in both these simulations, the shallow circulation was considerably deeper than that observed in the east Pacific, these papers support the connection between a surface temperature gradient and a secondary, low-level circulation embedded within a convectively driven deep circulation.

Chiang et al. (2001) compared the linear response of the tropical atmosphere to elevated heating and surface temperature gradients. They found that surface temperatures dominate the forcing of meridional winds at low latitudes where the surface temperature gradients are large (such as in the east Pacific), which would be consistent with our hypothesis. However, the possibility that the SMC is driven by shallow convection must also be considered. Using a simpler linear model with Rayleigh damping and Newtonian cooling developed by Wu et al. (2000), Wu (2003) compared the circulations generated by deep heating profiles and shallow heating profiles. Shallow heating profiles were shown to be quite effective in generating zonal overturning circulations, extending both above (into the midtroposphere) and below (to the surface) the vertical limits of the heating.

Two recent studies have suggested that SMCs between deep convection and nonconvecting regions are at least in part due to yet another mechanism: the pressure gradients associated with cloud-top radiative cooling of the low-level stratocumulus in the nonconvecting region. Data-driven, regional simulations of the equatorial east Pacific by Wang et al. (2005) reproduced an SMC similar to the observed. They found that elimination of radiative cooling by low-level clouds south of the equator reduced the strength of, but did not significantly weaken, the SRF. This suggests that cooling by cloud-top radiation from stratocumulus clouds south of the equator amplifies, but does not cause, the SMC. Bretherton et al. (2005) also found a low-level return flow from convecting to nonconvecting regions, driven by enhanced radiative cooling at the top of the PBL in the nonconvecting region. However, these simulations had no SST gradient, and the Wang et al. (2005) results suggest that the strong SST gradient is the dominant mechanism for forcing the SMC.

The SMC has potentially important implications for climate in the east Pacific, and perhaps elsewhere. For the idealized simulations we present here, we find there is significant transport of water out of the ITCZ region in the shallow return flow, far more than what is transported in the upper branch. Furthermore, much of this water is recycled into the boundary layer, and may modulate the thickness of the stratocumulus-topped boundary layers away from the ITCZ. These boundary layers are believed to have enormous impacts on regional and global climate (Mitchell and Wallace 1992; Li and Philander 1996; Philander et al. 1996; Nigam 1997; Bergman and Hendon 2000; Wang et al. 2005).

Our hypothesis is that the shallow meridional circulation is driven by the pressure gradients that develop in response to the local surface temperature gradient when deep convection is suppressed or absent in the ITCZ region. Section 2 presents a simple analytical model of the SMC as a sea-breeze circulation driven by regional gradients of surface temperature and pressure. The numerical model, domain, and parameters used for idealized simulations of the Tropics are described in section 3. Section 4 describes and analyzes the dynamics of the SMC that appears as part of the idealized simulations. A water budget for the ITCZ region is presented in section 5, along with some examination of



FIG. 2. Diagram of the environment that drives the shallow circulation. The solid black line represents the lower surface (the ocean). The dash-dot line indicates the top of the boundary layer, which typically is higher in the warmer ascent region. The gray lines describe the direction of flow. The lapse rates  $\Gamma$  in each region are indicated: the boundary layer has the well-mixed lapse rate  $\Gamma_b$ ; the ascent region lapse rate is approximately that of a moist adiabat; in the subtropical descent regions, the atmosphere has a more stable environmental lapse rate  $\Gamma_c$ .

the effects on boundary layer stratocumulus. Conclusions and future work are discussed in section 6.

# 2. The shallow meridional circulation as a large-scale "sea breeze"

The term "sea breeze" generally applies to an onshore wind driven by the temperature gradient between a warm land surface adjacent to a cooler ocean. The onshore sea breeze is accompanied by a compensating flow from land to sea, just above the boundary layer (Stull 1988; Wallace and Hobbs 2006). However, the same principle can apply to any low-level overturning circulation driven by surface gradients of pressure and temperature. In the east Pacific, where the SMC was first observed, there is an unusually large meridional gradient of SST maintained by regional air–sea dynamics (Mitchell and Wallace 1992; Philander et al. 1996; Raymond et al. 2004).

In the region of low-level convergence, the temperature in the boundary layer is warm and the surface pressure is low. To the north and south, temperatures are cooler and surface pressures are higher. Meridional flow at the surface is driven by the pressure gradient toward the convergence region. Above the boundary layer, the warmer temperatures create greater thickness between layers of equal pressure, and the pressure gradient reverses. These points are illustrated in Fig. 2. We approximate the lapse rate in the well-mixed boundary layer as one slightly less than a dry adiabat. The depth of the boundary layer is allowed to vary, typically being deeper over the warmer ocean. Above the boundary layer in the ascent region, frequent convection drives the environment toward a moist radiative-convective equilibrium, and the lapse rate is close to a moist adiabatic profile. In the descent regions, the atmosphere is typically more stable with a smaller lapse rate (especially at low levels, where there can even be an inversion).

In an isothermal atmosphere, pressure decays exponentially with height. For an atmosphere with a constant lapse rate  $\Gamma$ , the following useful formula can be obtained,

$$p(z) = p_0 \left(\frac{T_0 - \Gamma z}{T_0}\right)^{g/R\Gamma}, \qquad (2.1)$$

where  $p_0$  and  $T_0$  are the pressure and temperature at some reference level, g is the gravitational acceleration, and R is the dry gas constant. (2.1) can be used to calculate the pressure as a function of height for piecewise constant lapse rates above a specified initial condition  $p_0$ ,  $T_0$ . In this manner, we can calculate p(z) from the surface to the top of the boundary layer with one lapse rate ( $\Gamma_b$ ), and then for another lapse rate above the boundary layer ( $\Gamma_a$ ). At each column along the horizontal axis of Fig. 2, the pressure as a function of height is

$$p(z) = \begin{cases} p_0 \left(\frac{T_0 - \Gamma_b z}{T_0}\right)^{g/R\Gamma_b} & z \le z_b \\ p_b \left(\frac{T_b - \Gamma_a z}{T_b}\right)^{g/R\Gamma_a} & , \quad (2.2) \end{cases}$$

where  $p_b = p(z_b)$  is the pressure at the top of the boundary layer,  $T_b = T_0 - \Gamma_b z_b$  is the temperature at the top of the boundary layer, and  $\Gamma_a$  is the lapse rate above the boundary layer.

We use this simple model to try to answer two questions: First, are the temperature and pressure gradients observed in the east Pacific sufficient to drive a seabreeze-like circulation with its complementary return flow above the boundary layer? And, if so, at what altitudes should this return flow be found? For the purposes of illustration, we imagine that  $p_0$ ,  $T_0$ ,  $\Gamma(z)$ , and  $z_b$ vary linearly between specified values in the ascent and descent regions, which are 1000 km apart. As a first example, we set  $T_1 = 300$  K,  $p_1 = 1010$  hPa,  $z_{b1} = 2000$ m,  $T_2 = 295$  K,  $p_2 = 1013$  hPa,  $z_{b2} = 1000$  m, the boundary layer lapse rate  $\Gamma_b = 9.0 \times 10^{-3} \text{ K m}^{-1}$ , the moist lapse rate  $\Gamma_m = 0.006$  K m<sup>-1</sup>, and the environmental lapse rate  $\Gamma_e = 0.004$  K m<sup>-1</sup>. Thus for this example,  $\Gamma_a$  in (2.2) varies linearly from  $\Gamma_m$  to  $\Gamma_e$  between the ascent and descent regions. The surface pressures and temperatures are chosen to reflect values observed in the EPIC field program (McGauley et al. 2004; see Figs. 5 and 7). Given the linearly varying parameters, pressure and temperature profiles are computed for each column between the ascent and descent regions.



FIG. 3. (a) Perturbation pressure and (b) horizontal pressure gradient force computed from the model of the shallow circulation as a sea-breeze circulation, for the base values of each parameter. For the horizontal pressure gradient force, the zero contour is shown as the thick line. Contour intervals for pressure are 0.25 hPa, and for pressure force are  $1.0 \times 10^{-4}$  m s<sup>-2</sup>.

Figure 3 shows contours of perturbation pressure and the horizontal pressure gradient force,  $-(1/\rho)(\partial p/\partial y)$  (hereafter HPGF) in the (y, z) plane. There is a section of higher pressure in the ascent region, causing positive pressure gradients from z = 1.5 to z = 4.5 km, with a maximum near z = 3 km.

Since the parameters are made to vary linearly between the two ends of the domain, it is tempting to think that the pressure field would also have a constant gradient between them. However, this is not the case, due to the nonlinearity of (2.2), particularly in regards to the variation of the exponent which contains  $\Gamma_a$ . Looking at Fig. 3, we can see that the pressure gradient varies in the meridional direction, even reversing, so as to make the meridional extent of the shallow return



FIG. 4. As in Fig. 3, but for HPGF calculated for some variations on the parameters of the sea-breeze model: (a)  $T_1 = 301$  K, (b)  $T_1 = 299$  K, (c)  $p_1 = 1011$  hPa, (d)  $p_1 = 1009$  hPa.

flow finite in length. The zero contour is indicated in the figures, identifying the region of positive values of HPGF which would drive the SRF out of the convergence region.

The sea-breeze model shows that the SRF is very sensitive to the differences in pressure and temperature between the ascent and descent zones. For example, increasing the SST in the convergence zone by only 1° makes the return flow of the SMC 42% deeper, 58% stronger (as measured by the maximum HPGF), and

49% longer (Fig. 4a; Table 1). The depth, strength, and length of this and the other SRFs generated by the sea-breeze model are presented in Table 1. Decreasing this SST by 1° has a similar effect of shrinking and weakening the return flow by even larger amounts (Fig. 4b; Table 1). Equal changes in the surface temperature in the descent region cause almost exactly equal and opposite changes to the return flow region (not shown). Changes in the surface pressure in each region have similarly strong effects as changing the surface tem-

TABLE 1. Altitude, strength, and extent of idealized shallow return flow.

Т <sub>1</sub> (К)	p <sub>1</sub> (hPa)	z <sub>b1</sub> (km)	<i>T</i> <sub>2</sub> (K)	p <sub>2</sub> (hPa)	z <sub>b2</sub> (km)	$\Gamma_m$ (K m <sup>-1</sup> )	$\Gamma_e$ (K m <sup>-1</sup> )	z <sub>max</sub> (km)	z <sub>bot</sub> (km)	z <sub>top</sub> (km)	$\begin{array}{c} {\rm Max~HPGF}\times \\ 10^4~{\rm m~s^{-2}} \end{array}$	Extent (km)
300	1010	2000	295	1013	1000	0.006	0.004	2.94	1.50	4.75	1.17	570
301								3.41	1.27	5.87	2.19	850
299								2.47	1.84	3.39	0.30	190
	1011							2.92	1.01	5.25	1.96	1000
	1009							2.97	1.97	4.01	0.38	180
						0.0065	0.0035	2.79	1.50	4.28	1.19	550
						0.0055	0.0045	3.41	1.50	5.99	1.21	610

perature, as shown in Figs. 4c,d, and in Table 1. If the surface temperature  $T_1$  in the ascent region is cooler than 298 K, or if the surface pressure  $p_1$  falls below 1009 hPa, pressures remain higher at all altitudes above the descent region, and the SRF is eliminated. Generally speaking, the absolute values of these pressures and temperatures are not important, but it is the differences between the temperatures and pressures in the ascent and descent zones that control the reversal of the pressure gradient. The absolute values have only a very small effect, which come in through the dependences on the environmental values in (2.2).

It is the difference in the lapse rates, with the atmosphere cooling more quickly with height in the ascent region of our model that allows the pressure gradient to reverse a second time, placing a vertical limit on the SRF. Thus, bringing the lapse rates closer together, with  $\Gamma_m = 0.0055$  K m<sup>-1</sup> and  $\Gamma_e = 0.0045$  K m<sup>-1</sup>, increases the strength of the SRF only slightly, but makes  $z_{\text{top}}$  increase from 4.75 to 5.99 km. With the lapse rates further apart, with  $\Gamma_m = 0.0065$  K m<sup>-1</sup> and  $\Gamma_e = 0.0035$  K m<sup>-1</sup>, the depth is decreased to 4.28 km.

To summarize, our simple model shows that when sufficiently large and oppositely signed differences in surface pressure and temperature are maintained between two regions, the resulting reversal of the pressure gradients aloft act to drive the SRF from the warm region to the cold region. For parameters consistent with the observed SMC, this return flow is predicted to range from 2 to 5 km in altitude, and extend several hundred kilometers out of the warm region. The size and strength of the return flow is very sensitive to the differences in surface pressure and temperature, and somewhat sensitive to the differences in the lapse rates between the ITCZ and descent regions. In a later section, we will compare the simple model to the temperature and pressure fields generated by a full physics simulation.

### **3.** Three-dimensional simulations of an idealized ITCZ and shallow meridional circulation

#### a. Model configurations

The numerical model used for this study was version 2.0.2 of the Weather Research and Forecast Model (WRF). WRF is a regional, fully compressible model of the atmosphere suitable for simulations on a wide range of horizontal scales and grid resolutions (Michalakes et al. 2001; Skamarock et al. 2005). WRF uses high-order advection schemes on an Arakawa-C grid, with  $\eta = p_h/p_{hs}$  as a terrain-following vertical coordinate (although there is no terrain in the simulations presented here), where  $p_h$  and  $p_{hs}$  are the hydrostatic pressure and

the hydrostatic surface pressures, respectively (Laprise 1992). The time integration uses third-order Runge–Kutta time stepping (Wicker and Skamarock 2002).

Our intention is to construct the simplest geometrical environment that will allow the model to reproduce the essential features of the tropical meridional circulation, such as the ITCZ, the Hadley circulation, and the SMC. The model domain is a zonally periodic channel with free-slip walls on the north and south boundaries. The earth's curvature is neglected in the model geometry and all map factors are set equal to one. The Coriolis parameter  $f = 2\Omega \sin \phi$ , where  $\phi$  is the equivalent latitude. The southern boundary lies at the equator,  $\phi = 0$ , and the northern boundary lies at  $\phi = 30^{\circ}$ N, and the boundary conditions are that there is no meridional flow through these walls. The grid horizontal spacings are  $\Delta x = \Delta y = 20.87$  km, with 200 points in the zonal direction and 160 points in the meridional direction. In the vertical, 40 levels are used, with constant spacing in  $\eta$ ; this produces levels that are stretched in height, such that the lowest half-level (where values of u, v, and  $\theta$ reside) is at approximately 100 m, and the first 10 levels are below z = 2.3 km. The model upper boundary is defined by the hydrostatic pressure at the top of the domain, which for these simulations is  $p_h(top) = 4754$ Pa. There is no vertical flow across this pressure surface, though it does move up and down between 18 and 20 km altitude because of thermal expansion and contraction of the atmosphere below. The time step is 60 s.

The surface is defined to be an ocean with a fixed SST that varies only in the meridional direction. The SST is set to be 30°C at the equator, decreasing linearly to 20°C at the northern boundary,  $\phi = 30^{\circ}$ N. The associated gradients of sensible and latent heat fluxes are the only source of energy for the simulated tropical circulation.

This idealized SST distribution is quite different from the observed SST in the eastern Pacific, with its equatorial cold tongue and the SST maximum near 10°N. However, our goal in this investigation is to reproduce the SMC with the simplest model possible. We have placed the highest SSTs and their associated ITCZ at the equator. This choice eliminates the sensitivity of the meridional overturning to additional factors, such as the latitude of the ITCZ (Lindzen and Hou 1988), inertial instability (Tomas and Webster 1997), and easterly waves (Burpee 1975; Gu and Zhang 2002). In additional contrast to the observed circulation in the east Pacific, the simulation is designed to produce an SRF that moves northward from the equator; this serves to ease the dynamical interpretation for those of us confined to the Northern Hemisphere.

Parameterizations for unresolved physical processes

that are available with WRF 2.0 are used, all of which are closely related to commonly used schemes for research and numerical weather prediction. For microphysics, the WRF single-moment (WSM) five-class scheme of Hong et al. (2004) is used, which is derived from the prior five-class scheme of Hong et al. (1998). The sophisticated Rapid Radiative Transfer Model (RRTM) scheme is used for longwave radiation (Mlawer et al. 1997). Shortwave radiation is not included. Rather, the effects of upper-tropospheric heating and the generation of the tropopause is mimicked by the use of a relaxation scheme on the upper-level temperatures. Above z = 16 km, the atmosphere is relaxed to an isothermal profile with a relaxation time scale of 5 days. At each column in the model, the relaxation temperature is determined at each time step from the mean temperature in the levels above 16 km. This approach gives the model the freedom to determine the tropopause temperature, based on the intensity of the simulated convection, which is pushing into the upper troposphere. Despite the arbitrary choice of the altitude above which the relaxation is enforced, it also gives some freedom to the height of the tropopause. The tendencies for the longwave radiation and the upper-level relaxation are updated every 15 min.

There were two choices for each of the planetary boundary layer (PBL) and cumulus convection parameterizations that seemed appropriate for this study. For the boundary layer, these were the Yonsei University (YSU) scheme (Noh et al. 2003), which is an improvement of the Medium-Range Forecast Model scheme (Hong and Pan 1996), and the Mellor–Yamada–Janjic (MYJ) scheme (Janjic 1994), which is derived from the operational Eta Model. For cumulus convection, the updated version of the Kain–Fritsch scheme (Kain and Fritsch 1990; Kain 2004), which includes the effects of shallow convection, and the Grell ensemble scheme (Grell 1993; Grell et al. 1994; Grell and Devenyi 2002), were considered.

We found that the YSU PBL scheme<sup>1</sup> and the Grell ensemble cumulus scheme used together produced the most realistic SMC. This combination, along with the parameters described above, will be hereafter referred to as the control configuration. Results for these simulations and a brief comparison to the results of other combinations are presented in the next sections.

#### b. Results for the control simulation

The simulation begins with a resting atmosphere in hydrostatic balance, initialized with a mean tropical sounding (Jordan 1958). This initial state has no influence on the final state, except to determine the total mass of the dry atmosphere. As the simulation begins, the atmosphere responds to the strong gradient in SST by developing a meridional circulation. This circulation intensifies as the model begins to simulate deep convection near the equator. Between 15 and 30 days, the simulation begins to resemble the textbook tropical overturning. There is convection and strong vertical motion near the equator, a strong meridional flow away from the equator at high altitudes, a strong subtropical jet near 25°N, and a return flow back to the equator at low levels.

The simulation reaches a quasi-equilibrium state after 90 days. This state is shown in Fig. 5, in terms of zonally and temporally averaged meridional cross sections of the zonal, meridional, and vertical winds (u, v, v)and w), the potential temperature  $\theta$ , the relative humidity (RH), and the total moist heating. By moist heating, we mean the sum of the temperature tendencies caused by the microphysics and cumulus parameterizations schemes. In this figure, the temporal average is over model output every 3 h from days 91 to 120. Near the upper boundary, the atmosphere can be seen to be very stable, as a result of the relaxation to an isothermal profile above 16 km. Note, however, that the altitude of the stable layer or tropopause is higher near the equator, and then slopes downward to lower altitudes in the subtropics.

Along with the expected features of the deep circulation, the SMC is immediately evident. The SRF above the boundary layer resides between z = 1.8 and z = 4.0km, with maximum mean meridional velocities of 2.4 m s<sup>-1</sup>. Remarkably, the area of low-level flow out of the ITCZ region has the same triangular, descending appearance as the regions of positive HPGF created by the sea-breeze model in section 2 (e.g., Fig. 3a). A somewhat similar structure is suggested by the existing observations (see McGauley et al. 2004, their Fig. 5). Consistent with the existence of both deep and shallow overturnings, two peaks in vertical motion and diabatic heating, one near z = 3 km and another near z = 9 km, can be seen in Fig. 5. As we shall see below, these

<sup>&</sup>lt;sup>1</sup> After all simulations for this study were completed, the WRF model developers reported that the YSU PBL scheme in WRF version 2.0.2 (and some subsequent versions) had a bug that affected surface friction over the ocean. The surface drag coefficient was not allowed to increase with wind speed over the oceans, as is commonly done to account for increased surface roughness due to larger ocean waves. It was reported that changes in model results due to correction of this error were significant only in tropical cyclones. Since the surface winds in our simulations were rarely over 20 m s<sup>-1</sup>, which rarely occurred in and around the ITCZ, we can only assume that the effect of this bug on our results would be very small.



FIG. 5. Temporal (from 90 to 120 days) and zonal (over the entire domain) mean values of the fields generated by the control simulation: (a) u, (b) v, (c) w, (d)  $\theta$ , (e) total heating from resolved and parameterized convection, (f) relative humidity. In these and subsequent figures, dashed contours indicate negative values, and maximum values, minimum values, and contour intervals are indicated at the top of each plot.

modes of convection do not exist simultaneously in the same times and places.

Another, perhaps unexpected, feature in the mean flow of the control simulation is a significant northerly flow into the ITCZ region at higher altitudes, from z = 5 to z = 8 km, between the shallow and the deep return flows. As Fig. 5f shows, this midlevel inflow is very dry and thus reminiscent of the dry intrusions that have

been observed at midlevels in tropical west Pacific (Numaguti et al. 1995; Parsons et al. 1994; Mapes and Zuidema 1996; Yoneyama and Parsons 1999; Waugh and Polvani 2000; Cau et al. 2005), and now recently in the east Pacific (Zuidema et al. 2006). Most of these observed dry inflow regions occurred at lower altitudes than in our simulations, and were often caused by advection from the midlatitudes associated with synoptic-scale events. However, Takayabu et al. (2006) observed inflow layers into convective complexes in the western tropical Pacific, at altitudes similar to those in our idealized simulations. Whether or not our simulated dry inflow layer is equivalent to the observed dry inflow layers remains for future work.

#### c. Sensitivity to model physics and domain size

Since our focus in this paper is on the dynamics of the SMC, we will not offer a detailed analysis of the effects of varying model parameterizations. Instead, we briefly review the results of simulations with other model options. Figure 6 shows the mean meridional circulations generated by WRF simulations over days 91 to 120 using 1) the Grell cumulus scheme and the MYJ PBL scheme; 2) the Kain–Fritsch cumulus scheme and the YSU PBL scheme; and 3) the Kain–Fritsch cumulus scheme and the MYJ PBL scheme.

Result 1 is fairly similar to the control case, but the SMC is much weaker and a bit deeper. Result 2 has a much stronger shallow return flow, but it is deeper and higher than in the control case. Furthermore, without additional modifications, this simulation produced numerous tropical cyclones, which prevented the formation of a quasi-steady ITCZ and SMC. Tropical cyclogenesis was prevented by restricting the surface moisture flux computed by the PBL scheme to an upper limit of  $1.8 \times 10^{-4}$  kg m<sup>-2</sup> s<sup>-1</sup> (values above which were only seen in the cores of the tropical cyclones), and the more reasonable tropical circulation shown in the figure was produced. In case 3, the model did not appear to reach any kind of steady, zonal-mean circulation after 90 days, with or without limitations of the surface moisture flux. The reasons for this failure could not be determined, but it is clear that this arrangement of parameterizations would not be useful for this study.

Of the four combinations of PBL and cumulus parameterizations, the SMC in the control case bears the greatest resemblance to the EPIC observations. One unrealistic aspect of the control case is that it produces extremely low mean relative humidity in the subtropical descent region, with values less than 10% extending downward to less than 2 km above the surface for the region 1000 to 1500 km north of the equator. Along with the mean moist heating (Fig. 5e), this suggests that

(a) Mean V (m/s) max=5.30e+00 min=-5.72e+00 int=7.50e-01







6 4 2 500 1000 1500 2000 2500 3000 y (km)

FIG. 6. Temporal and zonal means of v for different choices of parameterizations: (a) Grell ensemble cumulus scheme with the MYJ PBL scheme, (b) Kain–Fritsch cumulus scheme with the YSU PBL scheme, (c) Kain–Fritsch with MYJ.

the boundary layer remains stratocumulus-topped with virtually no shallow cumulus convection until within a few hundred kilometers of the ITCZ, which is quite different from the boundary layer south of the east Pacific ITCZ. This, and the absence of heating from shortwave radiation, increases the low-level radiational cooling in the stratocumulus region, which could enhance the SRF, as suggested by Bretherton et al. (2005) and Wang et al. (2005).

Two additional simulations were conducted using the same parameterizations as the control case, but with smaller and larger domain sizes in the meridional direction. The same grid spacing and SST gradient were used, but with 140 meridional points for the smaller domain and 180 points for the larger domain. The resulting mean circulations were nearly identical to the control simulation, with a distinct SMC in each case (not shown). The only notable difference was that, with the larger domain, the SRF was a bit higher and deeper than in the control case, but this may have been due to the simulation not having reached a steady state by 90 days. Regardless, there does not appear to be any important sensitivity to domain size in the meridional direction.

### *d.* Variability of the simulated shallow meridional circulation

When zonally averaged over the entire domain, the circulation in the control simulation shows little temporal variability with no apparent pattern (not shown). However, on small scales, there is considerable local variability in the convection, which then causes variability in the surrounding wind fields; an illustration of these small-scale variations will be shown below.

Along with the small-scale variations that are caused by localized convection, there is a more persistent and larger scale variability of the flow. To show this, we first present Hovmöller diagrams of the u and v velocities averaged between 200 and 430 km north of the equator (model points 10-20), along with w and moist heating averaged from 0 to 168 km north of the equator (model points 1-8). Furthermore, these averages are restricted to only the middle quarter of the zonal length of the domain (grid points 75 to 125 out of 200), so that the larger-scale variability of the circulation is captured. Figure 7 shows these Hovmöller diagrams over a 10-day period. There is clearly a periodic variability of the flow in the boundary layer and in the shallow return flow. This modulation has a period of about 2.25 days. Upon comparison of the plots for u and v with those for w and moist heating, one can see that the periods of strong SRF are negatively correlated with convection in the ITCZ, while periods of strong boundary layer flow into

the ITCZ are positively correlated with convection. This observation supports our hypothesis that the SMC is weakened or eliminated when deep convection is present. The hypothesis is further supported by timelag correlation analyses between w at z = 9 km in the ITCZ region, moist heating at 9 km in the ITCZ, and vin the deep outflow region, versus v in the middle of the SRF, as shown in Fig. 8. Values for these variables were averaged over the same regions used to make the Hovmöller diagrams above. Convection and the SRF are highly anticorrelated, with peaks in convective activity leading minima in the strength of the SRF by just a few hours. Thus, the adjustment to convection, which suppresses the SRF, occurs very quickly. The anticorrelation of the SRF with v at 14 km, representing the outflow of the deep circulation, is even stronger.

The spatial evolution of this variability can be seen in Fig. 9, in terms of the v field at z = 2.2 km (where the SRF is quite strong), and the rain rate in the equatorial region, for four times, each 9 h apart. These figures also show the small-scale variations in convection and their influence on the local circulations, as mentioned above. At the first time, the ITCZ precipitation is more active in the western (left) side of the domain, and less active in the east. This region of enhanced precipitation propagates to the east (right) in the subsequent figures. Similarly, the SRF is seen to be weaker and greatly disrupted by areas of inflow in the more convectively active region, while the SRF is more robust in the less active region. These regions of suppressed/enhanced SRF propagate to the east in tandem with the enhanced/suppressed convection. As it turns out, this variability in the convergence, convection, and the SRF is associated with an eastward-propagating, wavenumber-1, convectively coupled wave. This wave, the dynamics of the idealized SMC, and its variability are discussed in the next section.

#### 4. Analysis

# a. An eastward-propagating, convectively coupled wave

Our simulation of the tropical circulation generates an eastward-propagating, convectively coupled wave. The wavelength matches the length of the domain (4180 km), with a mean period of 2.25 days, and a phase speed of 21.5 m s<sup>-1</sup>. The mean zonal flow in the simulated equatorial region, averaged from 0° to 10°N and from the surface to the simulated tropopause, is -1.75m s<sup>-1</sup>. Thus the phase speed of the wave relative to the mean flow is 23.25 m s<sup>-1</sup>. For shallow-water Kelvin waves, these speeds correspond to equivalent depths of



FIG. 7. Hovmöller diagrams for u, v, w, and total diabatic heating (cumulus and resolved) for days 70 to 80 of the control simulation, restricted to the middle quarter of the domain (grid points 75 to 125 out of 200). The (a) u and (b) v velocities are averaged between 1.9° and 3.8°N, where the shallow circulation is strong, while (c) vertical velocities and (d) moist heating are averaged between 0° and 1.4°N, to indicate convection in the ITCZ.

55 and 47 m for the Doppler-shifted and nonshifted phase speeds, respectively.

The essential structure of the eastward wave is shown in Fig. 10. These fields are perturbations from the mean state, composited over times when the strength of the SRF is one standard deviation below the mean in the middle quarter of the domain, as indicated in Fig. 7b. The data have been smoothed by repeated application of a 1–2–1 filter in both directions. The results show the horizontal structure of the propagating wave. The weak phase of the SMC is indeed correlated with the phase of the wave where there is maximum low-level zonal convergence. Also evident are significant meridional circulations not typically associated with a Kelvin wave (Matsuno 1966; Holton 2004). These circulations are generated by enhanced convection in the ITCZ region, which are indicated by the positive perturbation w and condensate fields in the middle levels (Figs. 10c,d). The fact that the strong vertical motions are restricted to the ITCZ along the equator, and do not smoothly follow the line of maximum low-level convergence that extends north from the equator (as in Fig. 10a) shows that the vertical motions are greatly enhanced by convection and are only marginally due to dynamical adjustment from the wave. Comparison of the low-level u and vfields (Figs. 10a,b) to those at z = 14 km (Figs. 10e,f) show that the wave tilts westward with height, such that the vertical structure is approximately that of a first baroclinic mode.

While not dominant in the real atmosphere, Kelvin waves of this frequency and scale are indeed detected in the Tropics (Wheeler and Kiladis 1999). The dominance of this particular wave in our simulation is obviously related to the maximum scale that the wave can achieve in this periodic domain. The much lower phase speed and equivalent depth, as compared to dry Kelvin waves (with speeds of 40–50 m s<sup>-1</sup>) suggests that this wave is indeed coupled to convection, and the close



FIG. 8. Time-lag correlation between the meridional velocity in the shallow return flow and either the moist heating in the ITCZ region (solid), vertical velocity in the ITCZ region (dashed), and meridional flow in the upper-level outflow (dash-dot). The heating and vertical velocity are at approximately z = 8.2 km, averaged from 0° to 1.6°N, and the meridional velocities are at approximately z = 2.2 km and 14.0 km, averaged from 1.9° to 3.8°N.

correlation of the wave to vertical motion and condensation along the equator certainly supports this. The vertical structure and the organization of the winds in the simulated wave are qualitatively similar to those observed (Wheeler et al. 2000; Straub and Kiladis 2003) and to those seen in the equatorial channel simulations of Kuang et al. (2005) and Peters and Bretherton (2006).

Fortuitously, this wave provides us with quasiperiodic variability in the simulated ITCZ and SMC. This variability, in turn, allows us to see the effects of increasing and decreasing convective activity on the SRF. In this sense, these eastward-propagating waves act in manner similar to the westward-propagating easterly waves in the real Tropics (Holton et al. 1971; Chang 1973; Burpee 1975; Gu and Zhang 2002; Serra and Houze 2002; Petersen et al. 2003), in that they modulate the convection in the ITCZ region. Furthermore, they do so in a manner that is even more advantageous to our study, in that dry Kelvin waves have no meridional circulation. Thus, the variations in the SMC associated with the different phases of the wave must be caused by variations in the convective activity, and not the wave itself.

# b. Comparison of temperature profiles and force balances with the sea-breeze model

In section 2 we argued that the essence of the SMC was the large pressure gradients at the surface, decreas-

ing with height at different rates due to the large temperature differences in the ITCZ and subtropical descent zones. How well does this compare with the SMC generated in the model? Fig. 11a shows mean vertical profiles of potential temperature at y = 30 km (in the ITCZ) and at y = 616 km (in the descent region). The essence of the sea-breeze argument appears to be confirmed: the temperatures at the surface are higher in the ITCZ region, but then cool faster with height, such that they reverse; the atmosphere is considerably more stable above the boundary layer in the descent region than in the ITCZ. This figure is quite similar to composites of observed profiles of potential temperature on the cold and warm sides of the oceanic temperature gradient presented by Pyatt et al. (2005, their Fig. 15).

However, the mean lapse rates  $(\partial T/\partial z)$  generated in the model, as shown in Fig. 11b, indicate that the simple model used in section 2 greatly oversimplifies the thermal structures involved. The only place where a constant lapse rate seems to be a reasonable approximation is in the lower levels of the ascent region (but above the boundary layer), where the lapse rate is approximately 5 K km<sup>-1</sup>; at all other levels in both regions, the lapse rate varies greatly with height. Nonetheless, there is a region of very high stability (and low lapse rates) above the boundary layer in the descent region. It is this warmer layer that allows the pressure to fall more slowly with height in the descent region, causing the pressure gradient to reverse a second time, and placing a finite depth on the SRF.

Figure 12 shows the mean HPGF (as in section 2) and the residual force left after summing the HPGF, Coriolis force, and the effects of vertical diffusion of meridional momentum as parameterized by the YSU PBL scheme (this includes all diffusive effects at every vertical level, including surface drag and entrainment). Both fields show a "nose" of positive force and residual acceleration, between z = 1.5 and z = 4 km, extending a few hundred km away from the ITCZ region. There is also a large region of positive residual force in the boundary layer, which serves to decelerate the lowlevel inflow as it approaches the equator; above the boundary layer, however, the air in the ITCZ starts with zero meridional momentum, and thus must accelerate poleward, creating the SRF. The HPGF above the boundary layer in Fig. 12a bears a reasonable similarity to the HPGF predicted by the sea-breeze model, as in Fig. 3b, but it is about 4 times stronger, and contracted to a distance about half as long. While the HPGF decreases to zero near y = 300 km, the peak flow speed of the SRF occurs closer to y = 500 km. Some additional acceleration can be seen in the area of positive values in



FIG. 9. Plan views of v at z = 2.2 km and the base 10 logarithm of the hourly rain rate in mm h<sup>-1</sup> (from both resolved and parameterized convection) in the control simulation. Here, 1.0 mm h<sup>-1</sup> is added to the rain rate to prevent logarithms of zero, and values less than 0.1 are suppressed. The dates in these figures (e.g., 04–01–06) are relative to the start of the simulation on the first day of January (01–01–00), with the last two digits referring to the hour. The green arrows show the motion of the region of enhanced convection. In this and subsequent figures of horizontally varying fields, the altitudes indicated refer to the mean height of the model level in hydrostatic pressure coordinates.

the residual force that appears in the northern (right) side of Fig. 12a, of which the Coriolis force makes the largest contribution.

### *c. Modulation of the shallow meridional circulation by deep convection*

Using the variability provided by the Kelvin wave to our advantage, we can examine the differences in the size and strength of the SMC when deep convection is active or suppressed. We computed the temporal and zonal means in same zonally limited region used to generate the Hovmöller diagrams in Fig. 7. The fields were separated into periods when v in the SRF was more than one standard deviation above or below average, and then separate composite means were computed for these two regimes.

Figure 13 shows the composites for v, w, and moist heating when the SRF is strong (left side) and when it is weak (right side). Comparing these means to those in Fig. 5, it is clear that when the SRF is stronger than average, there is less convective activity in the ITCZ; when the SRF is weak, there is much more convection in the ITCZ. In the weak SRF regime, peak values of moist heating and w in the ITCZ region are about 9 and 6 times larger, respectively, than in the strong SRF regime. The northerly flow in the boundary layer is sub-



FIG. 10. Horizontal structure of the Kelvin wave as shown by composites of deviations from the means of fields at three model levels, averaged over time intervals when the shallow circulation is 1 std dev below average in the region from x = 1568 to 2613 km: (a), (b) u and v in the boundary layer, (c), (d) vertical velocity and total condensate near 7.7 km, and (e), (f) u and v near 14.5 km. All fields have been smoothed repeatedly with a 1–2–1 filter.

stantially stronger and deeper, as is the upper-level outflow from the equator. Similar changes in the strength of the SRF, on individual days with active and suppressed convection in the EPIC region, were presented by Raymond et al. (2006, their Figs. 10 and 13).

The vertical profiles of heating by moist convection

in the overall mean, strong SRF, and weak SRF composites can be seen more clearly in Fig. 14a. This shows that the strong and weak SRF regimes are not differentiated by regimes in which deep convection dominates over shallow convection, or vice versa. Rather, both types of heating are present in both regimes, and the peak values of the deep heating are always higher



FIG. 11. Vertical profiles of (a) potential temperature and (b) lapse rate in the time and zonal mean state of the simulation, at y = 31 km (solid) and y = 616 km (dashed).

than the shallow heating. Although the ratio of deep to shallow heating increases as the flow evolves from the strong SRF to the mean state to the weak SRF, the changes are also marked by a large increase in convection of both types. The equivalent changes in longwave cooling in the ITCZ are trivial compared to the moist heating (not shown).

Certainly, shallow convection must play a role in the SMC: air cannot elevate an additional 1 to 3 km above the boundary layer without the small boost of latent heat release in the shallow cumulus clouds. We note, however, that the theoretical work of Wu (2003) indicates that a return flow driven by shallow heating alone would occur well above the maximum value of the shallow heating, and extend above the vertical limit of the heating. Rather, the peak meridional flow for the strong SRF and mean composites is near z = 2.5 km, which is approximately equal to the altitude of the peak shallow heating for the strong SRF composite, and below it for the mean (Fig. 14a).

It is the regional gradients of pressure and temperature that drive the low-level flow from the ITCZ back to the subtropical regions. When deep convection becomes widespread, this gradient is diminished or even reversed; and when deep convection diminishes, the pressure gradient is enhanced by the decreased heating rates, with even some cooling occurring between 4 and 6 km. The meridional variation of surface pressure is shown for the mean, strong SRF, and weak SRF composites in Fig. 14b. During the weak SRF (enhanced deep convection) composite, the pressure falls in the ITCZ region, slightly more so than it does away from equator; in the strong SRF composite, it rises. This correlation between the low-level pressure gradient and



FIG. 12. (a) Horizontal pressure gradient force and (b) residual acceleration in the shallow meridional circulation region for the zonal and temporal mean flow. The zero contour is shown as the thick line. Units are  $m s^{-2}$ . Forces are not computed at the first grid point along the southern boundary.



FIG. 13. Composite means over the middle quarter of the domain when the SRF is more than 1 std dev above or below mean strength: (a) v (m s<sup>-1</sup>), strong SRF; (b) v, weak SRF; (c) w, strong SRF; (d) w, weak SRF; (e) total moist heating (K s<sup>-1</sup>), strong SRF (note smaller contour interval) but same shading scale as in Fig. 5; and (f) total moist heating (K s<sup>-1</sup>), weak SRF, same contour interval as in Fig. 5.

the abundance of deep convection in the ITCZ is quite similar to that found in EPIC observations by de Szoeke and Bretherton (2005) and Raymond et al. (2006). As indicated by the sea-breeze model of section 2, small changes in the surface pressure difference between the ITCZ and descent zones can lead to a large change in the size and strength of the region of positive HPGF that drives the SRF. ITCZ moist heating, mean (solid), strong SRF (dash), weak SRF (dash-dot)





FIG. 14. (a) Vertical profiles of moist diabatic heating in the ITCZ region and (b) meridional profiles of surface pressure for the overall mean (solid lines), the strong SRF composite (dash lines), and the weak SRF composite (dash-dot lines).

### 5. Moisture transport in the shallow meridional circulation

#### a. Implications for the regional water vapor budget

Without knowledge of the SMC, nor the midlevel inflow, a water vapor budget for the ITCZ region would consist of the following components: inward transport of water in the boundary layer, additional fluxes from evaporation, outward transport at high altitudes as part of the deep overturning, and loss through precipitation in the ITCZ region. Clearly, this budget must be modified by the "interior" circulations—but are their contributions to the water budget significant?

Figure 15a shows vertical profiles of the mean meridional transport of water (vapor and all condensate) at various latitudes in the control simulation. These were computed from the time and zonal mean values of  $\rho vq_{\text{total}}$ , rather than the products of the means, such that









Mean Meridional Water Transport, 4N (solid), 6N (dashed), 8N (dash-dot)



FIG. 15. Vertical profiles of meridional transport of total water (vapor and condensate) at latitudes  $4^{\circ}N$  (solid lines),  $6^{\circ}N$  (dashed), and  $8^{\circ}N$  (dash–dot) for (a) the time-zonal mean, (b) the strong SRF composite, and (c) the weak SRF composite.

Budget region, divergence direction	Boundary layer inflow (kg m <sup>-1</sup> s <sup>-1</sup> )	Shallow return outflow (kg m <sup>-1</sup> s <sup>-1</sup> )	Midlevel inflow (kg m <sup>-1</sup> s <sup>-1</sup> )	Upper-level outflow (kg m <sup>-1</sup> s <sup>-1</sup> )	Surface evaporation (kg m <sup>-1</sup> s <sup>-1</sup> )	Precipitation (kg m <sup>-1</sup> s <sup>-1</sup> )	Residual (kg m <sup>-1</sup> s <sup>-1</sup> )
			Time zona	l mean			
Equator to 4°N							
meridional:	-123.5	22.5	-1.7	2.2	-45.3	145.0	-0.8
6°N	-87.0	8.7	-1.1	1.5	-71.1	147.2	-1.8
8°N	-59.9	2.1	-1.1	1.2	-93.7	148.7	-2.7
			Weak SRF c	composite			
Equator to 4°N							
meridional:	-154.1	8.4	-1.1	1.7	-46.0	223.0	-13.9
zonal:	-40.0	-7.9	2.0	0.1			
Equator to 6°N							
meridional:	-105.0	4.2	-0.7	1.5	-71.7	226.3	-7.2
zonal:	-54.7	-9.4	2.2	0.2			
Equator to 8°N							
meridional:	-66.4	1.4	-0.8	1.6	-94.2	228.1	-2.6
zonal:	-64.6	-10.3	2.3	0.3			
			Strong SRF of	composite			
Equator to 4°N							
meridional:	-100.7	49.1	-3.2	3.1	-43.5	62.0	25.0
zonal:	40.1	23.7	-4.7	-1.0			
Equator to 6°N							
meridional:	-66.2	18.4	-2.2	1.6	-69.3	62.9	19.4
zonal:	52.9	27.7	-5.3	-1.1			
Equator to 8°N							
meridional:	-45.4	4.8	-1.8	0.6	-91.5	63.7	10.5
zonal:	59.1	27.7	-5.6	-1.1			

TABLE 2. Meridional and vertical water transports in and out of the ITCZ.

eddy transports are included. The solid, dashed, and dash-dot curves shows the vertical profiles at the model grid points nearest  $4^\circ$ ,  $6^\circ$ , and  $8^\circ$ N latitude, respectively. It is immediately evident that transport by the SRF is a significant factor in the balance of water in the ITCZ region. In fact, there is far more water coming out of the ITCZ in the SRF than in the upper-level outflow associated with the deep overturning of the Hadley circulation. Of course, the inflow in the boundary layer is much larger still, and is largely balanced by loss due to precipitation. Also shown in Fig. 15 are the same profiles for the composite means of the weak SRF (when convective activity is high) and the strong SRF (when convective activity is suppressed).

The meridional water transport can be divided into four layers: the boundary layer inflow, the shallow return outflow, the midlevel inflow, and the upper-level outflow. The meridional flux was integrated vertically in each of these four sections; that is,

$$F_q = \int_{p_1}^{p_2} vq_{\text{total}} \, dp, \tag{5.1}$$

where the boundaries  $p_1, p_2$  of each section are defined by reversals of the sign of the flux  $vq_{total}$ . These water transports were calculated in pressure coordinates so as to be more consistent with the numerical grid of the WRF model: the hydrostatic pressures  $p_1$  and  $p_2$  are defined at the bottom and top of each computational grid box, whereas v and q reside at the center of each box. The integrated fluxes in each section and at each latitude are summarized in Table 2. Since the strong and weak SRF composites were computed from a subsection of the periodic domain, the fluxes in and out of the ends of the region must also be considered to close the water budget. For the strong and weak SRF composites, these are shown in the second row of data in Table 2. Also shown are the total evaporation and precipitation rates over the areas of interest, and the budget residuals.

A number of surprising results are contained in this table. For example, the water transport in the midlevel inflow is nearly equal to the water transport in the upper-level outflow. Although this midlevel inflow air is quite dry in the relative sense (see Fig. 5f), it still contains enough water to be a significant part of the upperlevel budget. In the meridional direction, much more water is transported out of the ITCZ at 4°N by the upper-level outflow when the SRF is strong, and deep convection is suppressed, than when deep convection is active (3.1 versus 1.7 kg m<sup>-1</sup> s<sup>-1</sup>). At 6°N, they are nearly equal (1.5 versus 1.6 kg m<sup>-1</sup> s<sup>-1</sup>), and at 8°N there is more meridional transport when deep convection is active (1.6 versus 0.6 kg m<sup>-1</sup> s<sup>-1</sup>). The more consistent distinction between the two regimes is that the meridional transport is nearly constant over this latitude range (around 1.6 kg m<sup>-1</sup> s<sup>-1</sup>) in the weak SRF composite, while in the strong SRF composite the transport decreases drastically with latitude (from 3.1 to 0.6 kg m<sup>-1</sup> s<sup>-1</sup>).

Furthermore, these numbers do not account for the upper-level zonal divergence present with the weak SRF composite and its associated Kelvin wave, as seen in Fig. 10, and vice versa for the strong SRF composite. When accounting for this divergence, as shown in the second row of data in the weak SRF composite section of Table 2, this disparity at 4°N is mitigated by a net zonal convergence of moisture in the strong SRF composite  $(1.0 \text{ kg m}^{-1} \text{ s}^{-1})$ , but nonetheless there is still less water transported out of the ITCZ at 4°N (per unit length) when deep convection is strong than when it is weak. The fact that less water is transported seems counterintuitive, but can be explained by two additional observations: first, that the midlevel inflow of water is larger when deep convection is weaker, thus increasing the water content of the upper-level ITCZ (see Fig. 13a); second, and much more significant, is that the upper-level outflow is at a much higher altitude (Fig. 13a). Therefore, more water is condensed and lost to precipitation before the air turns north and leaves the ITCZ, a result reminiscent of the claim that deeper convection leads to upper-tropospheric drying rather than moistening (Lindzen 1990; Lindzen et al. 2001).

Some of the humidity in the midlevel inflow appears to come from evaporating water that has precipitated out of the upper-level outflow. This can be seen from Fig. 13c and Fig. 13e, which show the vertical motion and the moist heating of the strong SRF composite. Between y = 400 and 600 km, there is a secondary region of positive moist heating and vertical motion from z = 10 to 14 km, with evaporative cooling occurring just below. This evaporative cooling indicates that precipitation is evaporating in the midlevel inflow. As can be seen in the composites of w (Fig. 13d) and v (Fig. 13a), this secondary heating lifts the upper-level outflow to higher altitudes. A similar, or even opposite, upper-level heating structure is not present in the weak SRF composite (even when smaller contour intervals are used).

While this change in the structure of the upper-level outflow appears as an interesting feature of the strong SRF regime, it is most likely a remnant of the deep convection associated with the weak SRF regime. The period of the wave, which modulates convection, is only 2.25 days. The moist convective outflow in the ITCZ moves away from the equator at about 8 m s<sup>-1</sup> or less. Thus, in the time that the SRF changes from weak to strong (1.13 days), the moist outflow from deep convective updrafts has only traveled about 700 km, or to near 6.3°N latitude, which is near where the secondary moist heating occurs.

Turning to the low-level flow, another feature that is evident in the profiles in Fig. 15, and is confirmed by the data in Table 2, is that water transport in the SRF decreases drastically as it travels away from the ITCZ. In the full zonal mean, the SRF moisture flux decreases from 22.5 to 8.7 kg m<sup>-1</sup> s<sup>-1</sup> between 4° and 6°N, and then from 8.7 to 2.1 kg m<sup>-1</sup> s<sup>-1</sup> between 6° and 8°N. Another view of this decline in the water flux can be seen in by comparing Fig. 5b and Fig. 5f, which shows that the SRF flows across a very strong gradient of RH. The wedge shape in RH seen in the lower left corner of Fig. 5f is also apparent in EPIC dropsonde data along 95°W, as shown by McGauley et al. (2004, see their Fig. 5c) and de Szoeke et al. (2005, see their Fig. 7b).

Where does this water in the SRF go? Does it precipitate to the surface? The total precipitation in each of these subregions is only 2.2 and 1.5 kg m<sup>-1</sup> s<sup>-1</sup>, respectively. Thus, precipitation cannot account for much of the loss. However, in these same subregions, the boundary layer inflow flux increases by far more than can be accounted for by the surface moisture flux. For example, in the zonal mean, the boundary layer influx increases by 36.5 kg m<sup>-1</sup> s<sup>-1</sup> between 6° and 4°N. However, there is only 15.8 kg  $m^{-1}$  s<sup>-1</sup> of surface evaporation between 6° and 4°N. It seems likely that the 13.8 kg  $m^{-1}$  s<sup>-1</sup> of moisture lost by the shallow return flow is contributing to this shortfall, either through evaporation of shallow precipitation, or mixing at the top of the boundary layer. The remaining 7.9 kg m<sup>-1</sup> s<sup>-1</sup> could be coming from evaporation of precipitation generated at even higher altitudes. The transfer of water from the SRF to the boundary layer is similar to the apparent transfer of water from the upper-level outflow to the midlevel inflow, with the distinction that it appears to occur to some degree at all times.

For the time–zonal mean over the entire domain, the residuals in the water budget are reasonably small, with values that are only a few percent of the larger terms in the budget at each latitude. For the strong and weak SRF composites, however, the residuals are considerably larger. They show that when the SRF is weak,



FIG. 16. Cloud water and negative longwave radiation temperature tendency in the shallow circulation region for the strong and weak SRF composites: (a) cloud water in kg kg<sup>-1</sup>, strong SRF; (b) longwave cooling tendency in K s<sup>-1</sup>, strong SRF; (c) cloud water, weak SRF; and (d) longwave cooling tendency, weak SRF.

water is accumulating in the atmosphere of the ITCZ region, and when the SRF is strong, water content decreases in the ITCZ region.

### b. Shallow return flow moisture and radiational cooling in the subtropical stratocumulus

Numerous studies have noted the significant impact on climate of the large, stratocumulus cloud decks over the colder waters of the eastern South Pacific, with emphasis on the high albedo they impart to that region (Mitchell and Wallace 1992; Li and Philander 1996; Philander et al. 1996). More recently, several studies have also considered the dynamic effects of these clouds through the strong radiative cooling they cause on the lower troposphere, not only through reflection of solar radiation, but also through enhanced emission of longwave radiation. Nigam (1997) argued that the longwave cooling from the stratocumulus clouds in the southeast Pacific is critical in development of strong southerly flow along the equatorial South America, which triggers the annual transition of the east Pacific climate from its warm phase to its cool phase. Looking more generally around the globe, Bergman and Hendon (2000) found that cloud radiative forcing contributed to about 20% of the tropical circulations. Using a fullphysics, regional model, Wang et al. (2005) showed more definitively that this enhanced cooling increases the strength of the low-level circulation between the east Pacific ITCZ and the southeast Pacific descent region.

In our idealized simulations, does the moisture carried in the SRF contribute to the development of shallow cumulus and radiational cooling in the descent region? We can attempt to answer this question by comparing the thickness of stratocumulus in the weak and strong SRF composites. Figure 16 shows close-up views of cloud water content and the temperature tendency due to the longwave radiation scheme (for plotting purposes, the negative of this field is shown). In the strong SRF regime, the cloud water concentration at the top of the boundary layer reaches a maximum value of 1.52  $\times$  $10^{-4}$  kg kg<sup>-1</sup>, while for the weak SRF, it is slightly greater, reaching just over  $1.60 \times 10^{-4}$  kg kg<sup>-1</sup> (the larger maximum value indicated at the top of the figure occurs along the equator). In addition, the depth of the cloud layer is slightly greater in the weak SRF regime. The maximum (negative) temperature tendency due to longwave radiation is also slightly higher,  $1.61 \times 10^{-4}$ K s<sup>-1</sup> (or 13.9 K day<sup>-1</sup>) in the weak SRF regime, versus  $1.58 \times 10^{-4}$  K s<sup>-1</sup> (13.7 K day<sup>-1</sup>) in the strong SRF regime. Consistent with its deeper cloud layer, the areal extent of greater cooling rates is larger in the weak SRF regime, and in fact, the column integrated cooling (weighted by density) below 5 km is about 10% greater than in the strong circulation regime (not shown). These results show that the stratocumulus boundary layer and its associated cooling are actually deeper and stronger when the SRF is weak, and thus the enhanced moisture transport of the strong SRF does not provide a positive feedback through increased cooling at the top of the boundary layer in the subtropics.

The enhanced cloudiness in the weak SRF composite is perhaps due to the additional moisture flux and evaporation in the boundary layer, as shown in Table 2. Nonetheless, we suspect that the water that descends from the SRF to the boundary layer inflow between latitudes 4° and 8°N (as discussed above) must influence the abundance and thickness of these low-level clouds. Unfortunately, because of vertical redistribution of water by the cumulus parameterizations, the vertical diffusion of the PBL scheme, the fall of precipitation, and various microphysical processes,<sup>2</sup> it is very difficult to trace back the sources of vapor and condensed water in the simulated stratocumulus layer. Furthermore, any firm conclusions regarding its radiative impacts would be inappropriate with these idealized simulations, due to the neglect of shortwave radiation. A detailed assessment of the effect of the SMC on sources and sinks of moisture in the stratocumulus boundary layer, and the resulting radiative feedbacks on the regional dynamics (as in Wang et al. 2005), remains for future work.

#### 6. Conclusions

This paper has presented a theory to explain the existence of the recently identified shallow meridional circulation (SMC) in the tropical east Pacific. The fundamental hypothesis is that the SMC is essentially a seabreeze-type circulation that develops in response to the strong SST gradients in the east Pacific, and is strongest in the absence of deep convection. This hypothesis has been supported in five ways. First, a review of earlier simulations showed that SMCs generally appear in conjunction with deep circulations when a significant surface temperature gradient is incorporated into the model (Schneider and Lindzen 1977; Schneider 1977; Grabowski et al. 2000; Larson and Hartmann 2003). Second, analysis with the sea-breeze model presented in section 2 shows that the observed surface pressure and temperature differences around the east Pacific ITCZ can lead to a reversal of the pressure gradient above the boundary layer which would drive the SRF. Third, our idealized, full-physics simulation, driven by the surface gradient of SST, reproduces an SMC quite similar to that observed in the east Pacific, and the structure of the pressure field that drives the SRF is similar to that predicted by the sea-breeze model. Fourth, the simulations show that the SRF is strongest away from deep convection, and is suppressed or even eliminated when strong, deep convection develops in the nearby ITCZ. Finally, the simulations show that the SRF is maximized below the altitude where Wu (2003) predicts an outflow circulation driven by shallow convection would appear.

Bretherton et al. (2005) suggested that SMCs from convecting to nonconvecting regions would be driven by the cloud-top radiational cooling in the nonconvecting region. The peak radiational cooling rates of about 13 K day<sup>-1</sup> in our simulated stratocumulus cloud deck is considerably higher than rates of about 5 K day<sup>-1</sup> estimated from observations (Bergman and Hendon 1998) and simulated in a regional model of the east Pacific (Wang et al. 2005). However, this peak value occurs in a very thin layer (essentially one model level), with more realistic values of 4-10 K day<sup>-1</sup> spread out above and below this peak value (Fig. 16). Furthermore, the simulation did not show large variations in radiative cooling in conjunction with large changes in the strength of the SRF. Still, it is possible that the part of the mean strength the SRF in our control simulation is due to this large longwave cooling in the stratocumulus zone.

The observations of the SMC in the east Pacific (Zhang et al. 2004) and the simulations presented here validate the very early predictions of such a phenomenon by Schneider and Lindzen (1977) and Schneider (1977). However, their findings also suggested that the shallow and deep meridional circulations would exist simultaneously in a steady state (see, e.g., Fig. 1 of Schneider 1977). This was because Schneider and Lindzen used a hydrostatic model with a fixed stratifi-

<sup>&</sup>lt;sup>2</sup> Unfortunately, changes in water species due to microphysics were not available as output from WRF version 2.0.2.

cation, with constant heat sources at upper levels of the troposphere to mimic the effects of globally averaged moist convection in the Tropics. Thus, their upper-level circulation (driven by a prescribed upper-level heating) and the low-level circulation (driven by the surface temperature gradient) were decoupled. In our model, and in reality, these circulations become entangled by deep convective updrafts. Deep convection suppresses, and can even reverse, the SRF by changing the lowlevel temperature and pressure fields. Such changes were seen in EPIC observations by de Szoeke and Bretherton (2005) and Raymond et al. (2006). While the deep and shallow circulations appear together in temporal and zonal means, they do not exist simultaneously at any single location. Rather, as indicated in Fig. 9, on a very local scale (20-100 km) the meridional flow evolves between the two extremes described by the strong SRF (weak convection) or weak SRF (strong convection) regimes. These two flow regimes are illustrated in Fig. 17.

Our success in reproducing the SMC is probably because of our use of a particularly strong SST gradient near the equator. Over most of the planet, the meridional profile of SST is nearly flat near the equator, and decays slowly with latitude. The linear profile of SST was chosen as the simplest possible profile for our study; fortunately, the large temperature gradient near the equator caused by this choice was not too far from reality in regards to the east Pacific, because of the very large SST gradient south of the ITCZ (McGauley et al. 2004; de Szoeke and Bretherton 2005). Furthermore, placing the SST maximum at the equator eliminated the possible effects of easterly waves, inertial instability, and the dynamic sensitivity of the strength of the deep circulation to the latitude of the heating maximum. Our choice of replacing shortwave radiation in favor of an idealized upper-level tropopause forcing may have led to some unrealistic aspects of the simulations, such as the very low RH values in the descent region, high radiative cooling rates in the stratocumulus cloud decks, and a possibly stronger-than-intended SMC.

A careful examination of the water transports associated with the mean inflow and outflow layers provided some interesting results. The largest meridional transport of water out of the ITCZ occurs in the SRF. Given the substantially larger temperatures at lower altitudes, this result is obvious in hindsight; however, one should keep in mind that the very existence of the SMC was only confirmed in the last few years! Its role in modulating the water content, convection, and upper-level transport out of the ITCZ has received, to our knowledge, no previous attention. The water transport in the SRF decreases rapidly with latitude, and it ap-



FIG. 17. Idealized diagrams of the tropical meridional overturnings for the (a) strong and (b) weak SRF regimes. Solid arrows indicate direction and magnitude (by their thickness) of the transport of water. Dashed arrows indicate vertical transport by precipitation and possible evaporation.

pears that much of this water is recycled into the boundary layer inflow.

With these issues in mind, we plan to further explore the dynamics of the ITCZ and the SMC. For example, to determine the sensitivity of the strength of the SRF to the local SST gradient, we will use more gradual profiles of SST away from the equator; to determine the sensitivity of the SMC to the location of the ITCZ and to equatorially asymmetric dynamics, we will use domains that extend across the equator, with SST maxima off of the equator; and for more accurate analysis of the simulated stratocumulus regions, we will use a more realistic shortwave radiation scheme, and consider different boundary layer and microphysics parameterizations. As computational resources increase, and the flexibility of the WRF model continues to improve, nested simulations with cloud-resolving resolutions in the ITCZ region may also be possible in the near future.

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