Processes Controlling the Mean Tropical Pacific Precipitation Pattern. Part II: The SPCZ and the Southeast Pacific Dry Zone

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(Manuscript received 29 August 2006, in final form 4 April 2007)

ABSTRACT

The nature of the South Pacific convergence zone (SPCZ) is addressed by focusing on the dry (and cool) zone bounded by it and the coast of South America through numerical experiments. As shown in a companion paper, this dry zone is due, to a large extent, to orographically forced subsidence. Here it is shown that the northwestward expansion of this dry zone can be explained by advection of low moist static energy by the trade winds. These results provide an explanation of the geometry of the western edge of the dry zone and, therefore, of the eastern edge of the adjacent SPCZ. Sea surface temperature underneath the SPCZ is enhanced by relatively high near-surface humidity through evaporative processes, which feeds back into its organization. However, in this model, this feedback is not critical for the existence of the SPCZ. The subsidence associated with the ITCZ in the North Hemisphere negatively affects the precipitation rate in the SPCZ. It was also found that the sensitivity of the forced response is largest for peak orographic heights below 3000 m, which indicates that the exact representation of the Andes in numerical models might not be as critical as that of lower orography such as that in southern Africa.

1. Introduction

An intriguing feature in the climatological rainfall distribution in the tropical Pacific is the South Pacific convergence zone (SPCZ), which has a bandlike structure oriented from the western equatorial Pacific to the southeast, reaching approximately 30°S, 120°W (Fig. 1, top). The SPCZ is best developed during the austral summer months (December to February; see review by Vincent 1994). Alternatively, we can shift our attention to the region to the east of the SPCZ, which is persistently free of convective rainfall year-round, and which might be considered a more robust feature of the climatology of the South Pacific than the SPCZ (cf. Fig. 1, top and bottom). This “dry zone” (as we will hereafter refer to it) is an archetypal example of a trade wind regime (Riehl 1979) and is considered a critical region for global climate (e.g., Pierrehumbert 1995; Bony and Dufresne 2005).

A theory explaining the SPCZ is still lacking, although some advances have been made toward this goal. Kiladis et al. (1989) attempted to address the effects of the presence of South America and Australia on the SPCZ by removing them in a numerical model while retaining the underlying sea surface temperature (SST) distribution. Their null results highlight the importance of the SST distribution in determining the flow pattern, probably through the low-level thermally driven hydrostatic pressure gradients (Lindzen and Nigam 1987). DeWitt et al. (1996), using a linearized version of their atmospheric general circulation model (GCM), diagnosed the relative importance of the SST distribution and convective heating in determining the zonal asymmetries in low-level moisture convergence, and their results indicate that the linear contribution from the dynamical forcing by SST is small, yet significant, in the SPCZ. In a full GCM with prescribed SST, the resulting forced low-level convergence could be enough to organize convection (e.g., Kiladis et al. 1989). These results suggest that a theory for the SPCZ should also address how the SST distribution is established.

Other studies have provided some insights into the subtropical portion of the SPCZ and other subtropical convergence zones, by studying how the atmosphere responds to prescribed convective heat sources near the
equator [e.g., the Amazon (Figueroa et al. 1995) and the western Pacific (Webster 1981; Matthews et al. 1996; Kodama 1999)]. The common result from these studies is that the atmosphere responds with ascent to the east and poleward of the prescribed heat source, which might then organize convection into horizontally tilted structures akin to the SPCZ. These simulations provide evidence that atmospheric dynamics has a preference for organizing into such structures. However, there is no reason to believe that, in the specific case of the SPCZ, we can externally specify any heat source.

A different perspective arises from the experiments reported in the first part of this study (Takahashi and Battisti 2007, hereafter Part I), in which the presence of the Andes alone forced the ocean–atmosphere to organize into a configuration that strikingly resembles the observed configuration over the Pacific: a zone of suppressed deep convection west of South America (the model’s “dry zone”) bounded on its western edge by a horizontally tilted band of rainfall similar to the SPCZ (Fig. 2). This result suggests that the influence from the east might be an important factor shaping the east–west distribution of rainfall and SST in the south tropical Pacific, including the northwest–southeast tilt of the SPCZ.

In this study, a modeling approach similar to Part I will be used to elucidate the physical processes responsible for the organization of the SPCZ and the dry zone in the model, which might yield insights into how this works in nature. The approach consists of experimentation with an atmospheric GCM coupled to idealized representations of the ocean and with a dry atmospheric GCM forced by idealized diabatic forcings. In section 2, the control experiment and its results are described. Next, in section 3, an experiment with the dry model will be used to explore the dynamics of the orographically forced subsidence in the dry zone, and a mechanism based on low-level advection of dry, cool air will be proposed to explain the northwestward extension of the dry zone. In section 4, we address the importance of air–sea coupling in the organization of the SPCZ and the effect of the strength of the ITCZ in the Northern Hemisphere (NH) on the SPCZ/dry zone system. Finally, in section 5 we summarize our conclusions and present a discussion of their implications.

2. Model and control experiment

The model used in this study is the same as that in Part I. We will briefly describe the modeling setup; more details can be found in Part I. The atmospheric component is a T63 primitive equation model with seven vertical levels and simplified physical parameterization schemes (Molteni 2003). It is coupled to an ocean mixed layer between 40°S and 40°N, which is forced by the surface heat fluxes produced by the atmospheric model and by prescribed “Q fluxes” that act as a proxy for ocean heat transports. Unless otherwise noted, the Q fluxes are independent of longitude and are symmetric about the equator. Poleward of 40° latitude, the sea surface temperature (SST) is prescribed to a reference field, which is a function of latitude only and symmetric about the equator. The control (CTL) run analyzed is exactly the same as the one denoted as ANDES in Part I. It is nearly an aqua-planet run in
which the only land and orography are those corresponding to the Andes Mountains; everywhere else is ocean. Insolation at the top of the atmosphere is set to the equatorially symmetrized annual mean.

In Part I we showed that the presence of the Andes forces subsidence of dry air to the west, which lowers the SST by evaporation, leading to a suppression of convection. According to the theory of Xie (1996), the resulting north–south asymmetry should propagate westward indefinitely (in this model setup, at least) but, as we see in Fig. 2, the asymmetry weakens toward the west and a double-ITCZ is found west of the date line, albeit with a stronger NH branch. The dry zone features reduced SST and specific humidity (Fig. 3, top) and has a zonal extent of about 10000 km near the equator, and about 2000 km near 30°S.

Along the western edge of the dry zone, we see a band of rainfall similar to the SPCZ, although it is weaker than observed. The modeled SPCZ is also narrower and less tilted than the observed (cf. Figs. 1 and 2), but these distortions also occur in this model even when forced with realistic boundary conditions (i.e., SST, orography, seasonality, etc.). The model SPCZ is located slightly poleward of the axis of maximum SST, which is a robust feature in nature (Kiladis et al. 1989; Figs. 1 and 2), and in Fig. 3 (top), we see that the relation of the SPCZ to SST is more clearly depicted in terms of departures from the zonal mean (hereafter referred to as “eddy” component). The SPCZ appears collocated with the axis of maximum eddy SST, which suggests that air–sea interactions are important in organizing the SPCZ. However, the observed offset between the maximum surface convergence and rainfall (Kiladis et al. 1989) is not found in the model results (not shown), probably because the vertical resolution of the model is insufficient to capture the vertical structure of the low-level flow.

Kodama (1992, 1999) argued for the importance of the low-level flow in providing the moisture for rainfall in the SPCZ. He found that most of the horizontal moisture transport came from within the dry zone, but argued that moisture transport along the SPCZ from the equatorial region was also important. The latter involves a poleward low-level jet that recurves around the South Pacific subtropical high. Our model results also indicate a reduction in the equatorward component of the flow along the SPCZ, but not enough to completely close the anticyclonic gyre (Fig. 3, bottom), so that the moisture transport into the model SPCZ is mostly from the dry zone rather than from this jet. However, the moisture budget in the model SPCZ (eastward of 150°W, within the 3 mm day\(^{-1}\) contour in Fig. 3, bottom) exhibits a very close balance between rainfall and local evaporation. As shown in section 4b, this departure from the observations is associated with the weakness of the modeled SPCZ, which in turn is related to the excessive strength of the ITCZ in the NH.

The vorticity dynamics in the model dry zone, as in nature, is well described by Sverdrup balance (Rodwell and Hoskins 2001, hereafter RH01; Fig. 4):

\[
\beta v = \frac{\partial \omega}{\partial p},
\]

which links the subsidence and the equatorward flow. The magnitude of the low-level eddy specific humidity (\(-3\) g kg\(^{-1}\); Fig. 3, top) is similar to an observationally constrained estimate [the National Centers for Environmental Prediction (NCEP) reanalysis; not shown], while the model eddy SST is somewhat lower than that observed [NCEP optimum interpolation (OI) SST]. The width of the dry zone near 30°S is smaller (2000 km) than in the observations (4000 km), consistent with the excessively small tilt of the SPCZ (cf. Figs. 1 and 2). On the other hand, near the equator the width of the dry zone compares well to observations, extending al-

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Fig. 3. (top) Eddy (departure from the zonal mean) SST (contours, interval 0.5 K) and 925-mb specific humidity (shading, g kg\(^{-1}\)) in the CTL run. The 3 mm day\(^{-1}\) precipitation contour (gray line) is included for reference. (bottom) Precipitation rate (shading, mm day\(^{-1}\)) and 925-mb streamlines from the CTL run.
most to the date line in both cases. As shown in Part I, the subsidence rate in the dry region compares favorably with observational estimates. In the following section we will present some experiments intended to shed some light on the dynamics of the dry zone.

3. Dry zone dynamics

To assess to what extent the structure of the dry zone can be explained by dry atmospheric dynamics alone (i.e., by the purely mechanical response to the presence of the Andes), we performed an experiment (DRY) with moisture removed and with the physical parameterization schemes replaced by the idealized forcing from Held and Suarez (1994). Thus, the effects of deep convection, turbulent mixing, and radiative tendencies are represented by relaxation to a prescribed equilibrium temperature profile, which is independent of longitude and is symmetric about the equator. To force the flow associated with the dry zone, a mountain range was placed centered at 63.75°W and 20°S, roughly the location of the Andes, with a half-cosine bell shape with widths of 15° and 40° in the zonal and meridional directions, respectively, and a peak height of 4000 m.

The low-level response to the west of the mountains (Fig. 5) is consistent with the picture of RH01. The subsidence and equatorward flow are also closely linked through Sverdrup balance [Eq. (1), Fig. 4]. In the theory of RH01, the mechanical forcing of the Andes makes the equatorward flow descend along isentropic surfaces, which can be considered to be set by the zonal mean flow, although farther into the Tropics diabatic processes play a more important role (RH01; Part I).

a. Dependence on height of orography

It is natural to expect that the height of the orography will have a strong effect on the magnitude of the subsidence and low-level equatorward flow. It might further be expected that this dependence will be linear for small topographic heights. The Andes extend almost half the depth of the troposphere, however, and so it is likely that the atmospheric response does not scale linearly with the height of the Andes. This was confirmed by a series of runs similar to DRY but with peak orographic heights ranging from 125 m to 5 km. The results show that the magnitude of the response increases strongly and linearly for small peak heights, but the sensitivity becomes weaker for heights above 3 km (Fig. 6). This reduced sensitivity suggests that the representation of the mechanical forcing of the subsidence by the Andes in low-resolution climate models will likely be adequate. However, it also suggests that low mountain ranges, such as the ones in southern Africa, might have a significant effect on climate. On the other hand, the zonal and depth scales (discussed in more detail in the following sections) are insensitive to the peak orographic height.

b. Depth scale

The constraint that the air flows along fixed isentropic surfaces is expressed as

$$\omega = -\frac{1}{\theta^\prime} \frac{\partial \theta^\prime}{\partial p},$$

(2)
where \( \theta_v/\theta_p \) is the isentropic slope associated with the zonal mean flow. Since the flow is also in Sverdrup balance, we can combine (1) and (2) to obtain the characteristic depth of the descending/equatorward flow:

\[
\Delta p = \frac{f}{\beta} \frac{\theta_v}{\theta_p} = a \tan(\phi) \frac{\theta_v}{\theta_p},
\]

where \( a \) is the radius of the earth and \( \phi \) is latitude. In Fig. 7 we show a cross section along the axis of the greatest subsidence/equatorward flow (77° W). We can see that \( \Delta p \) (with the isentropic slope estimated from this figure around 30° S) very closely depicts the vertical scale of the subsiding flow, which becomes quite shallow as it approaches the equator. This is also true in observational estimates from NCEP reanalysis (not shown) and is further confirmation that the mechanism proposed by RH01 is at work.

c. Zonal scale of the forced subsidence/equatorward flow

In the DRY run, the significant impact of the mountains is confined to be within \(~2000\) km to the west of the mountains (Fig. 5). This is in sharp contrast to the width of the dry zone along the equator (Fig. 3), which extends westward as much as 10,000 km from the mountains. Hence, the purely dynamical response of the atmosphere to the mountains does not explain the northwestward expansion of the dry zone. However, the localized subsidence/equatorward flow around 30° S is closely reproduced (Figs. 5 and 4) by the DRY run, so it is of interest to understand what sets its zonal scale in this context. In this section we rule out, based on experimental results, some scalings for the zonal extent \( L \) of the descending/equatorward flow forced by the presence of the mountains. However, the determination of the correct scaling for \( L \) has eluded us and remains a puzzle.

First, it should be noted that \( L \) is insensitive to the width of the mountain range, which was verified with a run, otherwise identical to DRY, in which the width of the mountain range was doubled (not shown). In section 3a we also noted that \( L \) is insensitive to the height of the mountains.

As in the CTL run with the full model, in the DRY run \( L \) is \(~2000\) km, which is on the order of the Rossby deformation radius \((NH/\beta)\) calculated at 30° latitude using either the tropopause height or a density scale height for \( H \) and the buoyancy frequency \((N)\) corresponding to a typical period of 10 min. Taking the Rossby radius of deformation as the scaling for \( L \), however, implies that the latter would have a dependence on latitude as \((\sin\phi)^{-1}\), which is not seen in the model results. On the other hand, if we calculate the deformation radius as \( Nh/\beta \), where \( h \) is the height scale equivalent to (3), the radius becomes

\[
\frac{N}{\beta} \frac{\theta_v}{\theta_z},
\]

which does not depend strongly on latitude. The relevance of this scale for \( L \) was tested in an experiment similar to DRY except that the meridional and vertical gradients in potential temperature that characterize the idealized forcing of Held and Suarez (1994) were each reduced by a factor of 2. The model results presented changes in \( \theta_v \) and \( \theta_z \) in the subtropics consistent with those in the forcing and, hence, the isentropic slope
$\theta_z/\theta_e$ was unchanged. [The depth scale of the response of the flow to the mountains was also unchanged, providing further evidence of the validity of (3).] Equation (4) predicts that $L$ should be reduced by 30% (due to the $\theta_z^{1/2}$ dependence of $N$) in this experiment. However, $L$ did not change appreciably, which indicates that (4) is not the relevant scale for $L$.

Yet another process that could determine $L$ is dissipation arresting the westward propagation of the effect of the presence of the mountains, which in turn could be associated with long Rossby waves, allowing a steady state to be achieved (e.g., Gill 1980). In this case, $L$ would be determined by how far the waves can propagate within the characteristic time scale of the dissipation, that is, $L \approx c_g t_d$, where $c_g$ is the group velocity and $t_d$ is the damping time scale. Since the strongest damping in our dry atmospheric model is frictional damping near the surface, we tested this hypothesis by repeating an experiment similar to DRY but with the frictional time scale doubled in the region $40^\circ$–$15^\circ$S, $120^\circ$–$70^\circ$W, which should have led to a doubling of $L$ if this were the relevant process. However, the results showed no change in $L$ (not shown), disproving this hypothesis. Similar negative results were obtained with the low-level thermal damping.

Although westward wave propagation does not appear to be likely for explaining the zonal extent of the response of the flow to the mountains, we speculate that the equatorial Rossby radius of deformation ($\sqrt{NH/(2B)}$) might provide a scale for $L$. Unfortunately, since this radius depends on $\theta_z^{1/4} \Omega^{1/2}$, testing the relevance of this scale in the model would require strongly changing the rotation rate of earth or the vertical stratification, which would have undesired (and stronger) effects on other aspects of the general circulation. Thus, although this scaling is plausible, at this point we offer it only as a possibility.

Furthermore, since the subtropics is a transition region (between the Tropics and midlatitudes), it is also possible that $NH/f$ and $\sqrt{NH/(2B)}$ describe the part of the subsidence region closest and farthest from the equator, respectively, in which case the subsidence would not be described by a single zonal scale.

d. Northwestward extent

Since dry dynamics do not explain the east–west extent of the dry zone near the equator, it is possible that thermodynamic processes involving air–sea interactions play the leading role. In particular, in Fig. 3 we see that the extension of the dry zone from the region of subsidence follows along the streamlines of the low-level flow. We hypothesize that advection of cool, dry air by the trade winds extends the cooling of the sea surface toward the northwest, thereby expanding the dry zone westward.

To test this hypothesis, we performed an experiment (SINK) similar to the CTL (full physics, coupled to a mixed layer) but with no topography. A moisture sink was then added to the lowest model layer; by relaxing the specific humidity in a localized region toward 1 g kg$^{-1}$ on a time scale of 1 day. This moisture sink, centered at $25^\circ$S, $75^\circ$W and extending $5^\circ$ from this point, is a proxy for the drying of the boundary layer produced by the subsidence forced by the Andes, but without the Andes perturbing the flow field.

The result for this experiment features a dry and cool zone that extends toward the northwest of the moisture sink (Fig. 8), in support of the hypothesis. To the west of the dry zone, an SPCZ-like feature in the rainfall field is collocated with a region of high eddy SST. The organization of the convection is enhanced by air–sea interactions, as we will show in section 4a. It is interesting to note that, although a strong north–south asymmetry was established, there is still an ITCZ in the southeast Pacific. This is not unlike many comprehensive GCMs, in which a double ITCZ structure coexists with an SPCZ. In this particular model, the mechanical enhancement of the equatorward flow by the Andes (e.g., Fig. 5), which enhances the influence of the advection in the far eastern Pacific, appears to be necessary to eliminate this feature.

To provide further support for our interpretation of the results in terms of low-level advection of the dry zone, we use an idealized thermodynamical–advective model (Takahashi 2006), which consists of an atmospheric mixed layer (AML, with moist static energy $m$) coupled to an ocean mixed layer (OML, with temperature $T$). This model consists of the following equations for the AML and OML, respectively,

\[
\frac{\partial m}{\partial s} + M(m - m_o) = SHF + LHF - R, \tag{5}
\]

\[
F = SHF + LHF, \tag{6}
\]
where $s$ is a coordinate following a streamline of the low-level horizontal mass transport $U$ (with magnitude $U$); $M$ is the mass flux across the top of the AML that includes large-scale motions and vertical turbulent mixing. The free troposphere just above the AML is characterized by $m_w = c_p T_a + L_e q_a$, where $T_a$ and $q_a$ are the corresponding potential temperature and specific humidity, respectively. Here $F$ is a prescribed net downward radiative surface flux plus ocean heat transport convergence and $R$ is a prescribed, noninteractive radiative cooling rate in the AML. The latent and sensible surface heat fluxes (LHF and SHF, respectively) are formulated using bulk formulas (see Takahashi 2006 for details) as

$$\text{SHF} + \text{LHF} = C(m_0 - m),$$

where

$$m_0 = c_p T + L_e q_s(T)$$

is the moist static energy of the ocean surface and $C$ is the effective transfer coefficient with units of mass flux.

The distributions for the fields above the AML ($\theta_w$ and $q_w$), the horizontal flow and mass fluxes ($U$ and $M$), and the surface radiative fluxes plus the ocean heat transport convergence ($F$) are idealized versions of the observed tropical Pacific zonal mean climate (Takahashi 2006), although the low-level horizontal flow ($U$) was modified to approximately match the zonal mean of the SINK run. Equations (5) and (6) can then be solved for the boundary layer moist static energy ($m$) and the sea surface temperature ($T$) by integrating equatorward along streamlines that start in the subtropics, where we assume $U = 0$. The climate so determined is discussed by Takahashi (2006).

Here, we will be concerned with the eddy fields resulting from a perturbation in $M$ representing the subsidence forced by the Andes. Linearizing $q_s$ about a reference temperature, we can write the equations for the perturbation moist static energy and ocean temperature:

$$U \frac{\partial m^*}{\partial s} = -[M] m^* - M^*([m] + m^* - m_w),$$

$$T^* = \left( L_e \frac{\partial q_s}{\partial T} + c_p \right)^{-1} m^*,$$

where $[()]$ and $()^*$ indicate basic state and perturbation respectively (in this discussion, “eddy” and “perturbation” are synonymous). A localized perturbation $M^*$ will be imposed in representation of the subsidence mechanically driven by the mountains in the CTL run or of the moisture sink in the SINK run. Downstream from this perturbation, we set $M^* = 0$, so (9) becomes

$$\frac{\partial m^*}{\partial s} = - \frac{m^*}{D},$$

where

$$D = U/[M]$$

is a local spatial scale over which $m^*$ and $T^*$ decay.

The perturbation $M^*$ chosen has a cosine-bell distribution centered at $25^°S$, $75^°W$ and a width of $5^°$. The amplitude of the subsidence perturbation was chosen to be $10^{-2} \text{ kg m}^{-2} \text{s}^{-1}$, about twice as strong as seen in the CTL run, in order to produce a response similar to that of the SINK run. The need for a large amplitude probably reflects the importance of the drying effect of the subsidence above the boundary layer, which is not considered in the simple model. Since the model is linear, this only affects the amplitude of the response.

Figures 9 and 10 show the results of this model and from the SINK run and we see that the simple model successfully reproduces both the spatial scale (12) and the relative sizes of $m^*$ and $T^*$ (10), providing further support that the extension of the dry zone depends significantly on low-level advection and local air–sea interaction. The wider structure of the dry zone in the GCM can be explained by the diffusence seen in the low-level flow, which most likely arises from the flow response to the SST gradients (Lindzen and Nigam 1987) and to the mechanical forcing by the Andes.
4. The convergence zone

a. Local air–sea interaction

We have seen that an SPCZ-like structure forms along the western edge of the dry zone extending to the northwest from the region of forced subsidence. Deep convection is expected to be suppressed in the dry zone due to the low near-surface moist static energy and enhanced subsidence. However, we have also seen that the convection along the edge of the dry zone is organized into a band, which is collocated with high eddy SST (Fig. 3). This local eddy SST maximum would favor the occurrence of convection both by promoting higher low-level moist static energy and by setting up horizontal pressure gradients within the boundary layer that would drive low-level convergence (Lindzen and Nigam 1987). Therefore, the organization of the rainfall and high eddy SST band is likely to involve the interaction between the ocean and the atmosphere. In fact, in a run similar to the CTL except that the ocean is not interactive, the SPCZ did not form, except for a hint south of 20°S (Fig. 9 in Part I).

A less restrictive way of testing the importance of local air–sea interaction in the organization of the SPCZ/high eddy SST system is to assess the impact of reducing the strength of the coupling between the eddy SST and the atmospheric flow. As shown in Part I, the most important factor controlling SST in the dry zone is specific humidity through evaporation. So, an efficient way of reducing the strength of the interaction of the eddy features in SST and in the atmosphere is to eliminate the contribution of the eddy component of specific humidity to surface evaporation. Based on this idea, we performed an experiment (EVAP) identical to CTL, except that the surface latent heat flux was calculated in the model using the instantaneous zonal mean surface specific humidity instead of the local value. The actual distribution of humidity was not directly modified during this procedure.

As expected, the resulting eddy SST was strongly reduced (cf. Figs. 3 and 11, top), verifying that the influence of evaporation on the equilibrium heat budget of the ocean mixed layer is dominant in setting up the high eddy SST. However, the low moisture signature of the dry zone was not strongly changed (Fig. 11, top) and the SPCZ is still present along its western edge with similar magnitudes of rainfall but more weakly organized than in the CTL run (cf. Figs. 3 and 11, bottom). These results indicate that, although the eddy SST distribution is important for the organization of the SPCZ into a band, it is not essential for the formation of the dry zone (defined in terms of atmospheric properties alone) or, equivalently, of the eastern edge of the SPCZ. Hence, the oceanic response is best thought of as an important positive feedback on the organization of the SPCZ rather than being its fundamental cause.

b. Influence of the ITCZ

The processes affecting the east–west differences between the dry zone and the SPCZ described in this
study do not directly depend on the ITCZ being predominantly in the Northern Hemisphere, even though both are mainly a result of the forcing by the Andes (Part I). In fact, in the SINK experiment (Fig. 8), as well as in many coupled GCMs, the SPCZ coexists with a Southern Hemisphere ITCZ in the eastern Pacific.

Nonetheless, the subsidence in the SH associated with the ITCZ is likely to hinder convection in the SPCZ. This is suggested in the experiments of Part I, where a stronger SPCZ is associated with a weaker eastern Pacific ITCZ (cf. Figs. 10 and 12 in Part I). Less clear, however, is how the ITCZ influences the geometry of the dry zone. For instance, (12) indicates that the downstream scale for the dry zone might be unchanged if both the zonal mean subsidence rate and the low-level flow were to change proportionally.

To assess the influence of the eastern Pacific ITCZ on the SPCZ/dry zone system in a more controlled setting, we performed an experiment (WEAKITCZ) similar to the CTL, except that the eastern Pacific ITCZ was artificially weakened by multiplying by 0.75 whatever latent heating was produced by the deep convection scheme at each time step within the region 0°–20°N, 180°–80°W. As a result, the ITCZ was displaced equatorward by 2.5° (not shown). The maximum latent heating in the ITCZ was reduced to about half relative to the CTL run, and the zonally averaged mass overturning (i.e., the Hadley circulation) was weakened by 25%.

Suppressing convection in the ITCZ region in the WEAKITCZ experiment caused the precipitation in the SPCZ area to increase by around 30%, and the main convection band to be displaced by about 10° westward and 2° poleward, although its horizontal tilt was not changed (Fig. 12). The eddy SST underlying the SPCZ increased from around 0.5 to 1 K (not shown) in a positive feedback to the organization of the SPCZ. The poleward component of the low-level flow in the SPCZ was also significantly increased, and the contribution of the moisture convergence to precipitation increased to around 20%. Generally speaking, the model SPCZ was closer to the observed.

The westward and poleward displacement of the SPCZ or, equivalently, the expansion of the dry zone with the weakening of the Hadley circulation, suggests an enhancement of the subsidence in the dry zone, which is indeed the case immediately to the east of the SPCZ (Fig. 12, bottom). This is associated with an increase of the width of the subsidence of about 25% (cf. Fig. 3, bottom, and Fig. 12, top). On the other hand, closer to the coast the changes in the subsidence are weaker (Fig. 12, bottom). In an experiment with the dry version of the model (section 3) with no orography, the addition of heating similar to the SPCZ (not shown) results in enhanced subsidence to the southwest, while to the east the subsidence and the trade winds are weakened, which would lead to low-level moistening and higher SST in the coupled model. According to this, enhanced convection in the SPCZ should lead to shrinking of the dry zone instead of the observed expansion, which remains to be explained.

5. Conclusions and discussion

The present work is the second half of a study of the processes that determine the large-scale climatological mean pattern of rainfall in the tropical Pacific. In Part I, it was found that orographic forcing of the atmospheric flow by the Andes can explain to a large extent the tropical climate pattern. The main processes involved are depicted schematically in Fig. 13. The mechanical interaction of the subtropical westerlies with the Andes orography creates a subtropical anticyclone and dry subsidence to the west of the mountains (Rob-
well and Hoskins 2001), which reduces the SST through evaporation (Takahashi and Battisti 2007) and leads to an asymmetric state with an ITCZ only in the Northern Hemisphere.

Besides the north–south asymmetry, the results of the experiments in Part I also featured a zonal structure in the Southern Hemisphere remarkably similar to the observed SPCZ/southeast Pacific dry zone structure, which are the subject of the present study. Previous studies of the SPCZ have focused on the control from the west, that is, the effect of heat sources associated with deep convection in the western Pacific warm pool and the Maritime Continent. The present study provides an alternative view by addressing the possibility of control from the east. From this perspective, the emphasis is shifted from the SPCZ itself to the dry zone bounded by it and the west coast of South America.

In the present study we have shown that the location and shape of the SPCZ can be explained by processes acting from the east, which create a large wedgelike dry zone in the southeast Pacific where SST is decreased and convection is inhibited. The geometry of the dry zone in the deep Tropics is mainly set by low-level advection of low moist static energy by the trade winds. A simplistic zeroth-order interpretation is that the tilt associated with the SPCZ is given by the orientation of the streamlines of the southeasterly trade winds. The width of the subtropical part of the dry zone or, equivalently, the separation of the tip of the SPCZ is given by the orientation of the streamlines of the southeasterly trade winds. The width of the subtropical part of the dry zone or, equivalently, the separation of the tip of the SPCZ from the coast of South America, results from the mechanical response of the atmospheric flow to the presence of the Andes through dry dynamics, although the processes setting this width have not been determined.

As expected, the magnitude of the response increases monotonically with the height of the mountains. However, the sensitivity decreases markedly when the height exceeds a certain value, which in our experiments with our dry model was around 3000 m. This suggests that climate models should not be too sensitive to the representation of the height of the Andes, but also that the lower topography in southern Africa can play a significant role in determining the climate in the tropical Atlantic, as discussed below.

Although the ultimate reason for rainfall to organize into a band is not clear, the eastern boundary of the deep convection in the SPCZ is set by the edge of the dry zone, which is oriented southeast–northwest. Furthermore, it was found that the local feedback from eddy SST, probably associated with the local enhancement of low-level moist static energy and convergence, is important for the organization of the convection. The strength of this interaction is likely to be sensitive to the convective parameterization scheme employed and the representation of the planetary boundary layer. It was also found that an artificial weakening of the eastern Pacific ITCZ in the Northern Hemisphere leads to the strengthening of the SPCZ, bringing both the modeled ITCZ and the SPCZ closer to observations. This experiment also showed a slight expansion of the dry zone, which is presently unexplained.

Previous studies support the view of a “western control,” in which the SPCZ is a response to localized convective heating, anchored to the west Pacific by the SST distribution (e.g., Kodama 1999; Matthews et al. 1996). In this view, the SPCZ results from moist-dynamical feedbacks on the eastern edge of the localized heating. A critical assumption implicit in this view is that the SST distribution can be taken for granted. Although the convection in the western Pacific is indeed associated with a strong east–west asymmetry in SST immediately along the equator (within 5° latitude) through Bjerknes feedback (Bjerknes 1969; Dijkstra and Neelin 1995), the equatorial SST gradient is not sufficient to explain the spatial distribution of rainfall in the southern tropical Pacific. In contrast, in the paradigm of an “eastern control” there is no need for a localized heat source in the western Pacific and both the SPCZ and the underlying SST distribution are explained in terms of advection of dry, cool air from the east and local air–sea interactions.

Prescribing a heat source on the Amazon is perhaps more adequate for discussing the South Atlantic convergence zone (SACZ), since the Amazonian land surface will be warmer than the surface of the Atlantic during the summer (e.g., Figueroa et al. 1995). Furthermore, different geographical factors, such as the presence of the bulge of Africa and the smallness of African orography, might suggest that the mechanisms de-
scribed in our study are not as relevant as in the Pacific. However, our results suggest that low orography like the one in southern Africa would have a significant drying and cooling effect locally and might, therefore, have an impact on the Atlantic ITCZ and SACZ. Interestingly, the westward extent of the dry zones in both Pacific and Atlantic are similar, despite the large differences in basin widths, which suggests that perhaps the processes highlighted in our study might produce an SACZ independently of the presence of South America.

Acknowledgments. This study is part of the doctoral dissertation of KT. The authors thank C. S. Bretherton, I. M. Held, G. K. Vallis, J. M. Wallace, and three anonymous reviewers for useful comments and suggestions, and F. Molteni for the development of SPEEDY.

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