Convective Control of ENSO Simulated in MIROC

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ABSTRACT

The high sensitivity of the El Niño–Southern Oscillation (ENSO) to cumulus convection is examined by means of a series of climate simulations using an updated version of the Model for Interdisciplinary Research on Climate (MIROC), called MIROC5. Given that the preindustrial control run using MIROC5 shows a realistic ENSO, the integration is repeated with four different values of the parameter, λ, which affects the efficiency of the entrainment rate in cumuli. The ENSO amplitude is found to be proportional to \( \lambda^{-1} \) and to vary from 0.6 to 1.6 K.

A comparison of four experiments reveals the mechanisms for which the cumulus convections control behavior of ENSO in MIROC as follows. Efficient entrainment due to a large \( \lambda \) increases congestus clouds over the intertropical convergence zone (ITCZ) and reduces the vertical temperature gradient over the eastern Pacific, resulting in a wetter ITCZ and drier cold tongue via accelerated meridional circulation. The dry cold tongue then shifts the atmospheric responses to El Niño/La Niña westward, thereby reducing the effective Bjerknes feedback. The first half of these processes is identifiable in a companion set of atmosphere model experiments, but the difference in mean precipitation contrast is quite small. On one hand, the mean meridional precipitation contrast over the eastern Pacific is a relevant indicator of the ENSO amplitude in MIROC. On the other hand, the nonlinear feedback from ENSO affects the mean state, the latter therefore not regarded as a fundamental cause for different ENSO amplitudes.

1. Introduction

It is widely known that the El Niño–Southern Oscillation (ENSO) phenomenon is a dominant interannual variability of the coupled atmosphere–ocean system. ENSO has the greatest global climatic impact among various modes of variability in the climate system; therefore, understanding the mechanism and predictability of ENSO, which has been advanced by a number of studies over the last two decades (Neelin et al. 1998; Wang and Picaut 2004, and references therein), is still an important issue.

The essence of ENSO is an oscillatory mode that can be captured by a simple linear framework (Schopf and Suarez 1988; Battisti and Hirst 1989; Jin 1997; Weisberg and Wang 1997). Simple models for ENSO can even represent two different types of El Niño identified in observations (Fedorov and Philander 2001). In the simple theoretical framework, the characteristics of ENSO-like modes have been well investigated, irrespective of their oscillatory mechanism, and can be easily controlled by perturbing several parameters.

However, the feasibility of controlling the simulated ENSO-like variability does not exist in state-of-the-art climate models [i.e., coupled atmosphere–ocean general circulation models (CGCMs)] because they involve complicated feedback processes and are therefore realistic. ENSO simulations by current generation CGCMs are certainly better than those by first-generation CGCMs.
(Meehl et al. 2005; Achuta Rao and Sperber 2006). However, the ENSO amplitude in the 23 CGCMs participating in the Climate Model Intercomparison Project Phase 3 (CMIP3) still varies from about 0.2 to 1.9 K (Guilyardi et al. 2009a). This diversity of the ENSO amplitude is one of the major sources of uncertainty on how ENSO will change in future climate change scenarios summarized in the Intergovernmental Panel on Climate Change (IPCC) Fourth Assessment Report (AR4) [Solomon et al. (2007); see also recent review by Collins et al. (2010)]. Capotondi et al. (2006) compared nine CMIP3 models and showed that differences among CGCMs are large even though some biases such as a narrow meridional extent and a westward shift in the wind stress as well as sea surface temperature (SST) anomalies in association with ENSO are commonly found.

Despite the fact that the slow time scale of ENSO is attributed to the dynamics in the upper ocean, recent studies that adopted multimodel approaches have suggested the dominant role of the atmospheric component in the diverse ENSO simulations by CGCMs (Schneider 2002; Guilyardi et al. 2004). In particular, studies using IPCC-class models point out the leading role of the cumulus convection scheme in ENSO simulations in the following two different ways (e.g., Neale et al. 2008; Guilyardi et al. 2009b).

One is the cumulus momentum transport (CMT) that acts to shift the climatological trade winds eastward. The ENSO-related wind stress anomaly is also shifted eastward, which counteracts the common model bias and therefore contributes to the realistic ENSO (Wittenberg et al. 2006; Wu et al. 2007; Kim et al. 2008; Neale et al. 2008). The CMT may compensate for the well-known cold bias in the central equatorial Pacific in coarse-resolution ocean models (e.g., Shaffrey et al. 2009) and hence may not be so effective in high-resolution simulations.

The other process is related to the thermodynamics of the cumulus convection. It has been argued that the Arakawa and Schubert (1974)-type cumulus scheme tends to generate deep convection even in dry conditions in the free troposphere. This deficiency can be partially improved by introducing an empirical cumulus triggering that prevents the deep convection unless the tropospheric relative humidity (RH) exceeds a threshold (e.g., Emori et al. 2005). Wu et al. (2007) included the cumulus triggering function and CMT in Community Climate System Model version 3 (CCSM3) and found that both ENSO and the Madden–Julian oscillation (MJO) were better simulated. They argued that more coherent stochastic forcing associated with the MJO may be the key for controlling ENSO in their model. Neale et al. (2008) came to a similar conclusion; they introduced a “dilution” approximation, which represents entrainment in the updraft, and showed that it lengthens the too-short ENSO period in CCSM3 without dilution parameterization.

The above studies advocate the importance of convection over the western–central Pacific. However, other works suggest that clouds over the eastern Pacific are the controlling factor of the ENSO (Toniazzo et al. 2008). By comparing 12 CMIP3 models, Lloyd et al. (2009) showed that a surface shortwave feedback associated with clouds over the Niño-3 region is the primary cause of model divergence in terms of ENSO amplitude. This gives us hope for constructing a metric to constrain the model’s ENSO, which is however still difficult (cf. Sun et al. 2009).

In our modeling activity that develops and maintains the Model for Interdisciplinary Research on Climate (MIROC), we could not address the question on ENSO since the MIROC version 3 (MIROC3) used for the IPCC AR4 has considerably underestimated the ENSO signal both in strength and amplitude. Fortunately, the updated version of MIROC5, which will be used for the Fifth Assessment Report (AR5), is capable of simulating a much more realistic ENSO (Watanabe et al. 2010). Furthermore, we found during the tuning phase of MIROC5 that the ENSO property changes systematically when we perturb a single parameter in the convection scheme. It is this mechanism that we examine in the present study (i.e., the ENSO property controlled by cumulus convection).

While the high sensitivity of ENSO to the convection scheme is consistent with previous studies, the mechanism found for the convective control of ENSO in MIROC has not yet been identified in other CGCMs.

This paper is organized as follows. In section 2, the new version of MIROC and the parameter sensitivity experiments carried out in this study are described. In section 3, the simulated ENSO characteristics are examined. In particular, the dependence of the amplitude and spatial structure of ENSO on the parameter chosen to perturb the convective activity is extensively investigated. Differences in the mean state among the experiments are also documented in this section, and the relationship between them and the ENSO property is briefly discussed. The mechanisms of the ENSO amplitude change are then examined in section 4, where we perform additional experiments and analyze not only monthly means but also 6-hourly outputs for convective activity. Through these, we aim at obtaining a process-oriented understanding for the convective control of ENSO. Section 5 gives a summary and discussion.

2. Model and experiments

a. MIROC5i

MIROC5 was developed based on MIROC3 (K-1 Model Developers 2004), which employs a global spectral dynamical core and implements a standard physics package.
for the atmosphere. The ocean and sea ice models comprise the updated Center for Climate System Research Ocean Component Model (COCO) (Hasumi 2006). A land model that includes a river module is also coupled to the system. In MIROC5, many of the schemes have been replaced either by implementing recent ones, some of which have been newly developed in our group (see Watanabe et al. 2010 for details). In this study, we use an interim version of MIROC5, called MIROC5i, which contains several minor differences from the official MIROC5. These include an update of the Asselin time filter, code optimization, the number of sea ice categories, a small change in the river routing, the treatment of a multilayer volcanic emission, and implementation of the so-called COSP simulator requested in the Cloud Feedback Model Intercomparison Project (more information available online at http://cfmip.metoffice.com). These differences are not crucial for the present climate simulation in the tropics, so the results presented in this paper are also applicable to MIROC5. MIROC5i shares the same physics package with MIROC5, which incorporates a level-2.5 turbulence closure, prognostic cloud scheme, cloud microphysics, and a prognostic scheme for number concentrations of cloud droplets and ice crystals [see Watanabe et al. (2010) for the individual schemes]. In particular, replacement of the convection scheme from the prognostic Arakawa–Schubert scheme (Pan and Randall 1998, hereafter PR98) to a newly developed scheme (Chikira and Sugiyama 2010, hereafter CS10) was critical for the present work. In CS10, a discrete representation of multiple cloud types is retained following PR98, but a state-dependent entrainment profile is incorporated. The assumption between the entrainment rate $\varepsilon$ and the updraft velocity $w$ is adopted following Gregory (2001):

$$\frac{\partial w^2}{\partial z} = 2\alpha(1 - \lambda)B - \frac{w^2}{z_0}, \quad \varepsilon = \lambda \frac{aB}{w^2}, \quad (1)$$

where $B$ is the buoyancy of a cloud air parcel, $a$ the parameter representing the ratio of buoyancy force used to accelerate the mean updraft, and $z_0$ the scale height of the dissipation term. The first equation in (1) describes the budget for $w$, while the second expresses that a certain fraction of the buoyancy-generated energy is consumed by entrainment, which is factored by the nondimensional parameter $\lambda$. For a prescribed environment, the buoyancy energy for $w$ is reduced more by entrainment with larger values of $\lambda$ and hence the deep cumulus cloud can develop less frequently. The behavior of cumulus convections in a single column model and in an atmospheric general circulation model (AGCM) has been reported in CS10 and Chikira (2010), respectively.

1 This version is referred to as MIROC4.2 for internal use.

### b. Sensitivity experiments

The preindustrial control experiment using MIROC5 showed a remarkable improvement in the ENSO amplitude and structure compared with MIROC3 (Watanabe et al. 2010). The standard resolution for MIROC5 is T85L40 for the atmosphere and 0.5° by 1.0° for the ocean component. We found that the model’s mean state and variability were not seriously altered by reducing the horizontal resolution of the atmosphere (but were strongly dependent on the vertical resolution), so we decided to use a coarser resolution of T42L40 for MIROC5i in this study. This was solely due to the increased computational burden of the new physics package in MIROC5.

Four coupled experiments are performed: in each of the 85-year-long experiments, $\lambda$ takes a different value from 0.5 to 0.575 (Table 1). The first five years are discarded from the analysis. We also carry out a set of companion AGCM experiments with shorter lengths of integration, which will be described in section 4.

### c. Observational data

To validate the mean state as well as the ENSO amplitude and spectrum in MIROC5i, the observed SST data by Ishii et al. (2006) are used. The data provide monthly SST on a 1° × 1° grid from January 1945 to December 2006. In addition, we use the Climate Prediction Center Merged Analysis of Precipitation (CMAP) data (Xie and Arkin 1997) and atmospheric fields obtained from the 40-yr European Centre for Medium Range Weather Forecasts (ECMWF) Re-Analysis (ERA-40; Uppala et al. 2005). These data are both recompiled into the monthly long-term means, which are then compared with the model climatology to examine whether MIROC5i reproduces several important aspects of the mean states. The monthly anomalies for the period of 1979–2006 and 1957–2002 are also used for the CMAP and ERA-40, respectively.

### 3. ENSO and mean states in MIROC

#### a. ENSO

The ENSO characteristics include various aspects such as the amplitude, periodicity, seasonality, and spatial

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**Table 1. Summary of the experiments using MIROC5i: $\sigma_{\text{nino}}$ denotes std dev of the Niño-3 SST anomalies (K); observation: 0.91 K.**

<table>
<thead>
<tr>
<th>Expt</th>
<th>$\lambda$</th>
<th>Length of integration</th>
<th>$\sigma_{\text{nino}}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>L500</td>
<td>0.500</td>
<td>85</td>
<td>1.63</td>
</tr>
<tr>
<td>L525</td>
<td>0.525</td>
<td>85</td>
<td>0.97</td>
</tr>
<tr>
<td>L550</td>
<td>0.550</td>
<td>85</td>
<td>0.77</td>
</tr>
<tr>
<td>L575</td>
<td>0.575</td>
<td>85</td>
<td>0.61</td>
</tr>
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</table>
The amplitude is a primary measure of ENSO and is defined by the standard deviation of the Niño-3 (5°S–5°N, 90°–150°W) SST anomalies ($\sigma_{\text{nino}}$), which is compared among the four experiments (Table 1). The ENSO amplitude is largest in L500, where $\sigma_{\text{nino}} = 1.63$ K, and smallest in L575, which shows $\sigma_{\text{nino}} = 0.61$ K. It is evident that $\sigma_{\text{nino}}$ is proportional to $\lambda^{-1}$, indicating that ENSO tends to be suppressed when the cumulus entrainment becomes efficient. The difference in $\sigma_{\text{nino}}$ between L500 and L575 is found to be considerably large when referring to the variation in $\sigma_{\text{nino}}$ in the CMIP3 models (see our introduction; Guilyardi et al. 2009a).

The time series of the Niño-3 SST anomalies and their power spectra are shown in Fig. 1. All of the experiments reveal an irregular occurrence of either El Niño or La Niña during a span of 80 years. Asymmetry between strong El Niño and moderate La Niña, which is an observed feature of ENSO (Burgers and Stephenson 1999; An and Jin 2004), is only seen in the experiments with small $\lambda$. The power spectra on which the observed spectrum is imposed exhibit that ENSO is exaggerated in L500 and vice versa in L575. The realistic amplitude lies between L525 and L500 (also confirmed from $\sigma_{\text{nino}}$ in Table 1). That the stronger ENSO occurs at lower frequencies is less visible but still clear. In L500 and L525, a single peak of 6.2 and 4.2 years dominates, whereas an additional high-frequency peak appears at 3–4 years in L550 and L575.
The ENSO structure in each experiment is identified by linear regressions of the monthly anomalies on the Niño-3 SST time series. To remove the dependence on amplitude, the regressed anomalies are presented per 1 K of Niño-3 SST anomalies in the subsequent sections. The change in ENSO structure can be visualized by taking the deviation of these anomalies from the model ensemble mean (i.e., average of the four experiments).

Figure 2 shows the SST regression maps for the observations and four CGCM experiments. A common feature of the ENSO-related SST anomaly pattern is obtained: strong warming in the central–eastern equatorial Pacific, weak warming in the Indian and Atlantic Oceans, and weak cooling in the western tropical and North Pacific regions. At the same time, differences among the four experiments are also detected. For larger $\lambda$, that is, L550 and L575, the positive SST anomaly in the equatorial Pacific has a maximum around the date line in addition to the center in the Niño-3 region (Figs. 2d,e). The Indian Ocean warming in these experiments is stronger by about 0.5 K, although the linkage with the central equatorial Pacific maximum is not clear. Overall, the model captures the ENSO anomaly pattern well, but a discrepancy with observations is found in the western Pacific due to the equatorial warm anomaly extending too far to the west.

The structural change in ENSO is more conspicuous in the atmospheric variables. The regressed anomalies in precipitation and zonal wind stress ($\tau_x$) along the equator both reveal a systematic shift in their zonal profiles (Fig. 3). In L500, the positive precipitation anomaly spreads over the entire Pacific, and the zonal stress anomaly has a peak at around 150°W. These are shifted westward, confined to the western–central Pacific, as $\lambda$ becomes large. This change appears to be consistent with the change in the SST anomaly pattern shown in Fig. 2. The regressed $\tau_x$ profile in L525 is closest to the observational estimate, yet the corresponding precipitation lacks the negative anomalies over the western Pacific. The relationship between the westward shift of the maximum in $\tau_x$ anomalies and the appearance of the high-frequency ENSO is consistent with

![Figure 2](image-url)
simple model experiments performed by An and Wang (2000).

In this version of MIROC, the ENSO signature is tightly coupled with the thermocline variability; therefore, the positive ocean temperature anomaly in the eastern equatorial Pacific during El Niño accompanies the negative anomaly in the western Pacific subsurface (contours in Fig. 4) that moves along the thermocline and eventually brings the turnabout of the ENSO phase. However, differences between the experiments exist; in particular, the positive temperature anomaly in L500 (L575) is weaker (greater) than the ensemble average in the central Pacific from the surface to a depth of 150 m (shading in Fig. 4). The additional peaks in SST anomalies in L550 and in L575 (Fig. 2) are thus the surface manifestation of the positive deviations of subsurface temperature anomalies around the date line (Figs. 4c,d). In summary, the regression anomalies in Figs. 2–4 indicate that ENSO-related anomalies tend to have a shorter zonal scale when the cumulus entrainment is more efficient (i.e., larger $\lambda$).

To elaborate on the coupled feedbacks associated with ENSO, we calculate the so-called coupling strength and heat flux feedback parameters, $\mu$ and $\alpha$ (Guilyardi et al. 2009b). Specifically, the coupling strength parameter is defined by the regression slope of the Nin˜o-4 $\tau_x$ anomaly on the Nin˜o-3 SST anomaly, whereas the heat flux feedback is defined by a similar regression but for the net surface heat flux ($Q_{\text{net}}$) in the Nin˜o-3 region. They respectively measure the efficacy of oceanic driving of the atmosphere owing to zonal SST gradients and the damping effect on the SST anomaly due to radiative and turbulent heat fluxes. The wind stress anomalies are not necessarily largest in the Nin˜o-4 region in the CGCM (cf. Fig. 3b), so $\mu$ and $\alpha$ are evaluated as a function of longitude by using the regional averages of $\tau_x$ and $Q_{\text{net}}$ anomalies centered at a longitude where the averaging area has the extent same as the Niño-4 and Niño-3 regions, respectively.

The zonal profile for $\mu$ is similar to the $\tau_x$ regression (Fig. 5a). This is not surprising because calculating $\mu$ at each longitude is equivalent to the zonal smoothing of Fig. 3b. When we focus on the Nin˜o-4 region (arrow in Fig. 5a), the magnitudes of the coupling in the four experiments are within the range of observational estimates (gray shading). However, the peak is shifted westward and amplified for large $\lambda$, and the zonal extent is shortened instead. The inverse relationship between the ENSO amplitude and $\mu$ can clearly be seen (even in the Nin˜o-4 region), which indicates that $\mu$ is not a useful metric for the ENSO amplitude, consistent with the analysis of the CMIP3 models (Lloyd et al. 2009). On the other hand, the Bjerknes feedback—a critical process for the ENSO growth—is known to strengthen for wider basin lengths (e.g., Jin 1996). Therefore, the westward confinement of $\tau_x$ anomalies (or $\mu$) suggests a reduction in the effective Bjerknes feedback, which may explain why the ENSO amplitude is suppressed for large $\lambda$. The thermal damping coefficient $\alpha$ is strongest in the Nin˜o-3 region, but there are no systematic differences among the experiments except that they all underestimate the damping effect estimated from observations (Fig. 5b).

We also examined the seasonality of the ENSO amplitude and the presence/absence of the two types of El Niño. In short, ENSO in the four experiments does not show a significant difference in terms of the above aspects. The seasonal cycles in $\sigma_{\text{Niño}}$ showed a maximum around December–January, as observed, but accompanied a secondary peak in July–August. A simple analysis for detecting the two types of ENSO (cf. Guilyardi 2006) reveals that all of the experiments involve both types and thus
simulate a hybrid ENSO evolution. Recent studies suggest that the relative contribution of the two types of ENSO may vary in a future climate (Yeh et al. 2009), so the difference among the four experiments may be amplified when we perform the doubling CO2 experiments with different values of $\lambda$.

**b. Climatological states**

By referring to a linear perturbation theory, changes in ENSO properties have often been related to changes in the tropical Pacific mean state (e.g., An and Jin 2000). There is, indeed, no doubt that the Pacific mean state impacts the ENSO behavior, so we demonstrate the differences in climatological mean fields among the four MIROC experiments in this section.

The annual-mean climatology of SST and precipitation is first compared between observations and the model ensemble average (Fig. 6). The tropical mean (30°S–30°N) SST has similar values between observations and MIROC5\textit{i} (26°C and 25.4°C, respectively), but the area of warm SST greater than 28°C is confined to the equatorial western Pacific (Figs. 6a,b). This is probably coupled with less mean precipitation around the Maritime Continent (Figs. 6c,d). Owing to several minor differences in the model code and coarser resolution of the atmosphere model, the mean distribution in MIROC5\textit{i} is somewhat worse than MIROC5 (Watanabe et al. 2010). Nevertheless, the climatological states shown are as good as the current generation CGCMs. In particular, the presence of the South Pacific convergence zone and reduction of the erroneous double intertropical convergence zone (ITCZ) structure over the eastern Pacific are the noticeable improvements from MIROC3.

The changes in annual-mean climatology among the experiments are identified by the deviations of the mean SST and precipitation from the ensemble averages (Fig. 7). We only present them for the tropical Pacific since the deviations outside this region are much smaller. In L500, the mean SST is higher than the ensemble average in the western North Pacific, north of Australia, and in particular in the central–eastern equatorial Pacific (Fig. 7a). The mean precipitation in L500 decreases over the western Pacific (Figs. 6a,b).
equatorial Pacific and ITCZ whereas it increases over the central–eastern equatorial Pacific (Fig. 7e). The patterns of these deviations are roughly reversed in L575 (Figs. 7d,h) and the tendency of the mean state changes in the other two experiments is mild but systematically proportional to $\lambda$ (Figs. 7b,c,f,g).

In short, the Pacific mean state tends to be warm and has a weak contrast between wet (i.e., ITCZ) and dry (i.e., cold tongue) regions when $\lambda$ is small, and vice versa.

The warm eastern Pacific in L500 implies a weak zonal SST gradient, thereby apparently favoring weak ENSO (Fedorov and Philander 2001; Yeh et al. 2009). Indeed, mean $\tau_h$ (thermocline slope) is strongest (steepest) in L575 and weakest (gentlest) in L500 (not shown). This contradiction can again be explained by the altered efficiency of the Bjerknes feedback suggested in the previous section. A test with a simple theoretical model does indicate that a stronger zonal gradient in the mean

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**Fig. 5.** As in Fig. 3, but for parameters for (a) the coupling strength, $\lambda$, and (b) the heat flux feedback, $\alpha$, in each experiment. Positions of the conventional values of $\lambda$ and $\alpha$ using the Niño-4 and Niño-3 averages are indicated by gray arrows; shading denotes the range of the observational estimate by Guilyardi et al. (2009b).

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**Fig. 6.** Annual-mean climatology of the SST (°C) in (a) observations and (b) model ensemble average. (c),(d) As in (a),(b), but for precipitation (mm day$^{-1}$).
SST can coexist with a weaker ENSO if the effective Bjerknes feedback is reduced (see appendix).

It should be noted that the changes in precipitation may be greater than those in SST. Namely, the maximum SST deviations in L500 and L575 in the eastern Pacific are about 0.5 and −0.3 K, which are smaller than $\sigma_{\text{nino}}$. Their fractions to the mean values are only about ±2% (shading in Figs. 7a,d). The precipitation deviations in L500 and L575 in the equatorial Pacific are about 0.8 and −0.6 mm day$^{-1}$, which account for more than 50% and −40%, respectively, of the mean precipitation (shading in Figs. 7e,h).

The observed interdecadal change in ENSO amplitude is sometimes related to the change in the background mean state of the Pacific atmosphere–ocean system (An and Wang 2000). In view of the linear perturbation theory, eigenanalysis of a simple coupled model reveals that the change in either the mean SST or thermocline slope modifies the growth rate of the ENSO-like unstable mode (e.g., An and Jin 2000; Fedorov and Philander 2001). Choi et al. (2009) suggest that a higher mean SST in the eastern Pacific favors the ENSO mode getting more energy from the background state because of higher sensitivity of the atmosphere to the given SST anomaly. The relationship between the mean state difference shown in Fig. 7 and the ENSO amplitude (cf. Fig. 1) is consistent with their argument (i.e., warm eastern Pacific accompanies strong...
El Niño/La Niña in L500). More specifically, a warm mean state in the eastern Pacific enables generation of precipitation and, hence, wind stress anomalies there, which leads to a strong Bjerknes feedback (Fig. 3). This simple interpretation explains the mean state change owing to different parameter values that cause changes in the ENSO amplitude and structure. However, the background state can slowly change by interacting with ENSO in a nonlinear fashion (cf. Choi et al. 2009). For example, active ENSO cycles show asymmetry between strong El Niños and moderate La Niñas (Figs. 1a,c), which can potentially contribute to increasing the mean SST in the eastern Pacific. Therefore we reserve drawing the conclusion that the change in mean state controls ENSO until we look into the behavior of individual convective systems more deeply, which we discuss in the next section.

ENSO has been argued to have a connection with not only the annual-mean state but also the mean seasonal cycle. Observations show an anticorrelation between the ENSO and seasonal cycle amplitudes (Gu and Philander 1995). A similar relationship is identified in the model ensemble of CMIP3, which is interpreted as the stronger (weaker) seasonal cycle in a model leaving less (more) energy for ENSO growth (Guilyardi 2006). In Figs. 8a–c, we present the mean seasonal cycles of the observed and modeled precipitation over the eastern Pacific together with the difference between L500 and L575. As compared with some of the CMIP3 models, which still suffer from an artificial double ITCZ (Bellucci et al. 2010), the model ensemble mean is found to remarkably resemble the CMAP data in terms of the magnitude of the ITCZ in June–September and a weak double ITCZ structure during February and May (Figs. 8a,b). Furthermore, the ITCZ is stronger in L575, as it accompanies drier conditions to the south for June–September when ENSO begins growing (Fig. 8c). This difference indicates that the seasonal cycle in precipitation is stronger in L575 and weaker in L500, consistent with the arguments in previous studies.

The latitude–time cross section of SST also reveals a good coincidence between the model and observations (not shown). However, it was harder to reproduce the seasonal cycle along the equator (Figs. 8d,e). While the ensemble average SST shows a westward propagation, as in observations in the eastern Pacific, its phase lags by about two months and the propagation turns eastward in the western Pacific. The model overestimates (underestimates) the seasonal cycle of \( r_e \) in the western (eastern) Pacific, even though the annual-mean zonal profile is remarkably realistic (cf. Fig. 9a of Watanabe et al. 2010). In L575, the SST seasonal cycle in the eastern Pacific further lags that in L500 but is not necessarily stronger (Fig. 8f). It should be noted that the difference shown in Fig. 8f does not include the annual-mean component, which is much larger (Figs. 7a,d). In the Niño-3 region, for example, the annual-mean SST difference between L575 and L500 is \(-0.78 \) K compared to the June–August (JJA) difference of \(+0.16 \) K. Overall, the change in \( \lambda \) affects more the annual-mean states than the seasonal cycle.

The mean atmospheric fields—RH, vertical motion, and diabatic heating due to cumulus convections—over the eastern Pacific are presented in Figs. 9 and 10. Since the differences in these fields among four experiments are systematic, they are evaluated only in terms of the differences between L500 and L575, as shown in Fig. 8c.

The latitude–height section of the annual-mean RH in the model ensemble mean captures the primary features appearing in ERA-40: drying of the subtropical middle troposphere, a moist boundary layer to the south of 20°N, and a moist tropopause centered at \(-10^\circ\)N (Figs. 9a,b). The northern (southern) subtropics also have a moist (dry) bias in the model. The RH difference between L500 and L575 reveals that the middle troposphere is drier over the southern off-equator when \( \lambda \) is large (Fig. 9c). Secondary positive and negative differences are found over the northern subtropics and near the tropopause, respectively. A tendency analysis shows that the change in mean RH is attributed to a dynamical process, that is, adiabatic subsidence, but not to the thermodynamics associated with local convections (not shown). Indeed, differences in the vertical pressure velocity (\( \omega \)) over the Niño 3 and ITCZ regions (the latter defined as \( 5^\circ–12^\circ \)N, \( 90^\circ–150^\circ \)W) indicate that the meridional overturning circulation over the eastern Pacific is strengthened by large \( \lambda \) (Figs. 10a,b). A similar tendency is found in the zonal-mean circulation (not shown). Thus, the drier troposphere over the cold tongue in L575 is linked with less precipitation through enhanced subsidence (Fig. 7h).

The mean cumulus heating (denoted as \( Q_{\text{cum}} \)) has different vertical profiles over the Niño-3 and ITCZ regions (Fig. 10c). Over the cold tongue, characterized by a cool sea surface, deep cumuli rarely develop and thereby the heating has a single peak at \(-850\) hPa, corresponding to shallow cumuli. Over the ITCZ, such shallow heating is collocated with deep heating, which has a peak at 600 hPa and reaches 200 hPa. The cumulus detrainment profiles indicate that the model reproduces an observed trimodal structure of the shallow and deep cumuli and the middle-level cumulus congestus (Johnson et al. 1999). The differences in \( Q_{\text{cum}} \) demonstrate that deep heating is reduced over the cold tongue, while the heating is suppressed only in the upper troposphere over the ITCZ (Fig. 10d). The latter is reasonable because a larger \( \lambda \) reduces buoyancy in the cumulus plume and hence prevents the development of deep cumulus. A notable feature in Fig. 10d is that the suppressed deep heating in the ITCZ accompanies
enhanced heating below 600 hPa. This suggests that more cumulus congestus is released with larger $\lambda$, and the resulting vertically integrated $Q_{\text{cum}}$ becomes positive, which is consistent with the stronger mean ascent (Fig. 10b). A part of the positive $Q_{\text{cum}}$ is compensated for the evaporative cooling of raindrops, but we confirmed that $Q_{\text{cum}}$ prevails in the total tendency due to physical processes ($Q_{\text{phy}}$).

The large-scale vertical motion depends not only on the diabatic heating/cooling but on the vertical temperature gradient in the environment (cf. Pierrehumbert 1995). The temperature gradient as measured by $S = -\partial \theta / \partial p$,
where \( \theta \) is potential temperature, in the ensemble mean climatology shows a contrast between the stable free troposphere and the conditionally unstable near-surface layer below 850 hPa (Fig. 10e). The difference in \( S \) between L575 and L500 indicates that the temperature gradient is reduced within 250–650 hPa over the both regions (Fig. 10f). Since the mass flux is proportional to \( S^{-1} \), the change in stratification also accelerates the meridional circulation.

It is clear from Fig. 10 that the large \( \lambda \) results in intensified subsidence over the cold tongue, which is responsible for suppressing deep convection and the resultant precipitation. Three processes linking the entrainment of individual convective systems and the large-scale subsidence may be possible.

1) Efficient entrainment directly prevents deep clouds over the cold tongue, thereby intensifying subsidence.
2) Efficient entrainment stimulates congestus over the ITCZ, which intensifies ascending motion and thereby subsidence due to mass conservation.
3) Efficient entrainment reduces \( S \) in the middle-upper troposphere, which strengthens subsidence even without change in \( Q_{\text{phy}} \) over the cold tongue.

Analyses presented in the next section will clarify which is most suitable to explain the mean state change between L575 and L500. Note that the tropical atmosphere cannot maintain a strong horizontal temperature gradient, so 2) and 3) above may occur simultaneously due to suppressed deep cumulus over the ITCZ. There may also be the other indirect radiative processes associated with nonconvective clouds (e.g., Toniazzo et al. 2008), but we have not found a significant difference in low clouds and cloud–radiative forcing over the cold tongue region.

4. Mechanism for the convective control of ENSO

As argued in the previous section, concluding that the ENSO amplitude is attributed to changing mean states may not be relevant because of two reasons: it does not explain how and why the mean state is modified and it lacks a possible feedback from perturbations (i.e., ENSO) to the mean state. However, regardless of the interpretation, the results in section 3b indicate that the contrast between the ITCZ and cold tongue is inversely proportional to the ENSO amplitude and may be a good indicator of ENSO in MIROC5i.

To prove the above idea, we defined a cold tongue dryness (CTD) index as the normalized precipitation difference between the two regions (Fig. 11a):

\[
\text{CTD} = (P_I - P_N)/P,
\] (2)

Fig. 9. (a) Annual-mean climatology of RH averaged over 90°W–180° in ERA-40; contour interval is 10% and the light (dark) shading indicates RH less than 30% (greater than 80%). (b) As in (a), but for the model ensemble average. (c) Difference of the annual-mean climatological RH between L500 and L575 (latter minus former); contour interval 1%.
where $P_I$ and $P_N$ are the annual-mean precipitation climatology over the ITCZ and Niño-3 regions and $\mathcal{P} = (P_I + P_N)/2$ is their average, respectively. The higher the CTD index is, the stronger the meridional precipitation (or moisture) contrast exists in the mean state. The CTD index has the upper limit of CTD = 2, where $P_I/\mathcal{P}$. Figure 11b shows the scatterplot of $\sigma_{\text{nino}}$ against the CTD index in observations (cross) and CGCM (red dots). The ENSO amplitude is clearly shown to be inversely proportional to the CTD, and the $\sigma_{\text{nino}}$–CTD relationship in the observation lies between L525 and L550.

Given that the CTD measures the ENSO amplitude in MIROC5i well, we address the question of what is the cause of the mean state change. According to a close relationship between the mean precipitation and RH (especially at ~600 hPa) identified in Figs. 7 and 9, the composite precipitation with respect to the 600-hPa RH ($\text{RH}_{600}$), together with the histogram of $\text{RH}_{600}$, is calculated using monthly data (Fig. 12). In all of the experiments, the mean $\text{RH}_{600}$ is about 20% (75%) and the probability distribution is positively (negatively) skewed over the Niño-3 region (ITCZ) (Fig. 12a). Again, there is a systematic change among the four experiments: a low (high) RH$_{600}$ occurs less (more) frequently over the cold tongue in L500, and vice versa in L575 (thick curves). The tendency for change is opposite over the ITCZ (thin curves). The composite precipitation indicates that the change occurs in a wet condition, where RH$_{600}$ is roughly higher than 60% (Fig. 12b). In L500, heavier precipitation is found over the Niño-3 region, while it is suppressed in the ITCZ. Both panels in Fig. 12 are consistent with the change in annual-mean states, but the implications are different. The difference in composite precipitation suggests the alternation of convective processes in a similar environment, that is, the direct effect of $\lambda$, and the difference in the RH$_{600}$ histogram may indicate that the different strength and frequency of ENSO affect the environmental conditions for convection, that is, rectification to the mean state.

To isolate direct effects of convection from coupled feedback, we repeated the experiments using the atmospheric component model (i.e., AGCM) of MIROC5i.

Fig. 10. (a) Ensemble and annual mean climatology of the vertical motion ($\omega$) averaged over the Niño-3 region (gray) and the ITCZ area (black). (b) Difference of the annual-mean climatological $\omega$ between L500 and L575. (c),(d) As in (a),(b), but for the cumulus heating. (e),(f) As in (a),(b), but for the vertical temperature gradient $S$. 

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The SST and sea ice were prescribed by the monthly climatology taken from ensemble averages of the CGCM runs. In addition to these AGCM runs (L500a, L525a, L550a, and L575a), four AGCM runs (L500b, L500c, L575b, and L575c) were carried out to examine whether convection in the ITCZ or over the cold tongue drives the meridional moisture contrast. Since the AGCM lacks the atmosphere–ocean interaction that generates large interannual variability, a 5-yr integration was found to be sufficient for comparing the mean states.

The CTD indices from the eight AGCM experiments are imposed on Fig. 11 (blue dots, green and purple asterisks). To facilitate a comparison with the coupled results, we used $\sigma_{\text{nino}}$ in the corresponding CGCM run. As in the CGCM, the CTD is found to be larger for larger $\lambda$, but the difference is fairly small. This indicates that the coupled feedback greatly amplifies the mean state differences and, thus, the mean state as represented by the CTD partly results from ENSO. The dependence of CTD on $\lambda$ in the AGCM corresponds to the difference in composite precipitation (Fig. 12b), whereas the difference in CTD between the set of CGCM and AGCM runs (e.g., L500 and L500a) is reflected in the difference in probability occurrence for RH$_{600}$ (Fig. 12a). The effect of the air–sea coupling works to reduce the mean precipitation contrast in any case because of asymmetry in the atmospheric response to El Niño and La Niña (tail of the RH$_{600}$ histogram can lengthen only during El Niño).
We found that the difference between the CGCM and AGCM in Fig. 11 is reduced when the CTD is simply defined as $P_I - P_N$ (not shown), which indicates that the air–sea coupling also increases the mean precipitation over the eastern Pacific as a whole.

It is worth noting that the CTD values in L500b and L575b were small and close to the value in L500a, while those in L500c and L575c were near the value in L575a. Regardless of the background value of $\lambda$, L500b and L575b use $\lambda = 0.5$ over the ITCZ region and $\lambda = 0.575$ over the Niño-3 region, and vice versa in L500c and L575c (cf. Table 2). The contrast between L500b/L575b and L500c/L575c clearly suggests that the modified entrainment efficiency affects the deep cumulus regime, which then changes the precipitation over the subsidence region via meridional circulation (e.g., Fig. 10). Among the three possibilities listed previously (end of section 3b), processes 2) or 3) are thus likely but not process 1).

The question why changes in convection in the ITCZ drive other changes over the eastern Pacific is addressed by calculating the composite of $Q_{\text{cum}}$ with respect to the convective available potential energy (CAPE) using 6-hourly outputs of the AGCM (only for the first two years). Shown in Fig. 13 are the composite $Q_{\text{cum}}$ (contour), its deviation from the ensemble average (shading), and the change in frequency of occurrence for CAPE (curve with dots) using the ocean grids between 15°S and 15°N. Unlike in Fig. 11, this kind of regime composite diagrams in the CGCM looks nearly identical to that of the AGCM (not shown). As in the time mean state (Fig. 10c), the composite $Q_{\text{cum}}$ reveals three major peaks identified for CAPE greater than 1500 m$^2$ s$^{-2}$: boundary layer cooling, low-level heating, and upper-tropospheric deep heating. When compared with the ensemble average, deep heating is intensified while shallow heating is weakened in L500, and vice versa in L575. These changes occur for large CAPE. The fraction of CAPE greater than 2500 m$^2$ s$^{-2}$ in the convective environment (i.e., nonzero CAPE) exceeds 46% in the ITCZ region but reaches only 8% in the Niño-3 region.

The mechanism for how the entrainment parameter $\lambda$ affects characteristics of convection and the atmospheric circulation over the eastern Pacific is summarized in Fig. 14, which schematically illustrates the conditions in L500 and L575. When the entrainment is not efficient (small $\lambda$; Fig. 14a), convection in the ITCZ is dominated by the deep vertical scale. The meridional circulation in the upper troposphere may be intensified by this deep diabatic heating, but be partly canceled by the large vertical temperature gradient ($S$). Therefore, subsidence cannot prevent intermittent occurrences of deep convection over the cold tongue region, leading to a reduction in the meridional precipitation gradient. In contrast, a too efficient entrainment (large $\lambda$; Fig. 14b) suppresses deep convection and stimulates congestus in the ITCZ, which is balanced by strong ascent in the lower troposphere.

**Table 2. Summary of the AGCM experiments.**

<table>
<thead>
<tr>
<th>Expt</th>
<th>$\lambda$</th>
<th>Length of integration</th>
<th>Remarks</th>
</tr>
</thead>
<tbody>
<tr>
<td>L500a</td>
<td>0.500</td>
<td>5</td>
<td>$\lambda = 0.575$ over the Niño-3 region</td>
</tr>
<tr>
<td>L525a</td>
<td>0.525</td>
<td>5</td>
<td></td>
</tr>
<tr>
<td>L550a</td>
<td>0.550</td>
<td>5</td>
<td></td>
</tr>
<tr>
<td>L575a</td>
<td>0.575</td>
<td>5</td>
<td></td>
</tr>
<tr>
<td>L500b</td>
<td>0.500</td>
<td>5</td>
<td>$\lambda = 0.575$ over the Niño-3 region</td>
</tr>
<tr>
<td>L500c</td>
<td>0.500</td>
<td>5</td>
<td>$\lambda = 0.575$ over the ITCZ region</td>
</tr>
<tr>
<td>L575b</td>
<td>0.575</td>
<td>5</td>
<td>$\lambda = 0.500$ over the ITCZ region</td>
</tr>
<tr>
<td>L575c</td>
<td>0.575</td>
<td>5</td>
<td>$\lambda = 0.500$ over the Niño-3 region</td>
</tr>
</tbody>
</table>

Fig. 12. (a) Histogram of the monthly RH600 over the Niño-3 region (thick curves) and ITCZ (thin curves): color convention as in Fig. 3. (b) Composite of the cumulus precipitation with respect to RH600 over the two regions.
intensifies subsidence over the cold tongue. These processes set a stronger dynamical constraint that prevents deep convection over the cold tongue and therefore results in a sharp meridional precipitation contrast. Since either L525 or L550 was closest to observations in many respects, the reality will be between the two states shown in Fig. 14. The differences between the two states, which can be measured by the CTD, are then amplified by the atmosphere–ocean coupling (Fig. 11) with which the mean state and ENSO cooperatively modify their structure and intensity (cf. section 3 and appendix).

5. Summary and discussion

Given that the preindustrial control run by the new version of the climate model MIROC (MIROC5) shows not only realistic ENSO but also high sensitivity of the ENSO amplitude to the detail in the cumulus convection scheme, mechanisms for the possible convective control of ENSO were investigated using an interim version of MIROC5, referred to as MIROC5i. Four experiments were carried out with changing values of the parameter $\lambda$ that alters the efficiency of entrainment rate in cumuli. They reveal that the ENSO amplitude is proportional to $\lambda^{-1}$ and varies from 0.6 to 1.6 K.

The mechanisms of the efficient entrainment process reducing the ENSO amplitude, and vice versa, are summarized as follows.

(i) Efficient entrainment due to large $\lambda$ suppresses deep cumulus and increases congestus over the ITCZ region while reducing $S$ over the eastern Pacific. Together they accelerate the meridional circulation (Figs. 10 and 13).

(ii) Enhanced subsidence over the cold tongue region prevents the intermittent occurrence of deep convection, leading to a strong meridional precipitation contrast in the eastern Pacific (Figs. 7 and 12).

(iii) A dry atmosphere over the cold tongue favors westward displacement of the ENSO-related precipitation and wind stress anomalies, which weaken the effective Bjerknes feedback (Figs. 3, 4, and A1).

(iv) While the interannual variability associated with ENSO always acts to reduce the mean meridional precipitation contrast, the effect is less obvious with weak ENSO, and therefore the precipitation contrast is maintained (Fig. 11).

In the above sequence of processes, (iii) and (iv) represent the nonlinear feedback between ENSO and the mean state and are highly efficient when the ENSO is strong (e.g., L500 in Fig. 11). The other two are the atmospheric processes and thus identifiable in an AGCM even though the AGCM constrained by the prescribed SST prevents the atmosphere from being sensitive to the change in entrainment rate.

The results presented in this study, especially those shown in Fig. 11, have several implications.
hand, the mean state (i.e., CTD) gives a good measure of the ENSO amplitude simulated in MIROC. In a paleo-ENSO study based on CGCM sensitivity experiments, Merkel et al. (2010) found a clear anticorrelation between the ENSO amplitude and meridional contrast over the eastern Pacific. This agrees well with our results and suggests the robustness of the meridional contrast as a measure of ENSO regardless of the processes that set the mean state. On the other hand, change in the mean state should not be interpreted as the cause of different ENSO amplitudes in the presence of strong interaction between the mean state and ENSO. Since the atmosphere–ocean coupling, or the ENSO–mean state interaction, alters the mean state to a specific direction, using the AGCM to tune and determine a suitable set of parameters for the CGCM is not relevant.

There has been controversy over whether the radiative–convective equilibrium is driven by the ascending or descending branch in the tropics. Recent studies suggest that, contrary to intuition, the subtropics where the radiative cooling balances the adiabatic heating is the driver, not the equatorial ascent associated with the deep convection (e.g., Pierrehumbert 1995). The results of the present study are apparently inconsistent with the above idea, but not necessarily so. Our parameter experiments are a type of model ensemble in which neither radiative forcing nor processes are perturbed. When the convective process was perturbed, the deep cumulus cloud regime was found to be sensitive to the change in $\lambda$ and thus triggers the change in the tropical climate. However, we should bear in mind that cloud fraction changes in response to the changes in SST and circulation. We examined the cloud–radiative feedback in MIROC5i, which had a minor role but will largely be model dependent.

Although the results are not shown, we found that the equatorial waves in the atmosphere also show some dependence on $\lambda$. Both Kelvin and Rossby waves have shallow equivalent depths for larger $\lambda$, consistent with the change in the vertical structure of $Q_{\text{cum}}$. A stronger MJO signal is identified with a smaller $\lambda$ in the frequency–wavenumber spectra. This agrees with the argument that a better representation of stochastic forcing improves ENSO as reported in previous works (Wu et al. 2007; Neale et al. 2008), even though it will not be critical in MIROC5i.

The meridional precipitation contrast defined by the CTD also measures the erroneous double ITCZ structure in the model. The close relationship between the CTD and the convective process is consistent with the arguments made by Bellucci et al. (2010), who showed that the double ITCZ error is related to the misrepresentation of cumulus cloud regimes. However, a model that reproduces a realistic single ITCZ may not necessarily show a better ENSO (e.g., MIROC3) because multiple errors often compensate for each other in CGCMs. Instead, we anticipate that the ENSO amplitude varies with the magnitude of errors for the ITCZ in a model.

We chose $\lambda$ as the control parameter because the entrainment profile was known to be tightly coupled with the convective mass flux as evaluated from cloud-resolving simulations (Lin 1999), although it can hardly be observed and thus contributes to a large uncertainty in GCMs.
However, the convection scheme may involve other parameters that potentially influence ENSO. For processes that affect the coupled cloud–circulation fields over the eastern Pacific, parameters in other schemes such as for cloud physics and turbulence may also be candidates that control ENSO. A systematic exploration of the ENSO control mechanisms in a CGCM requires a physics ensemble that has been extensively used for investigating the model’s climate sensitivity. Such computations are expensive, so needs to be done in future studies (Toniazzo et al. 2008; Philip et al. 2010).

As described in the introduction, individual CGCMs may have different causes of errors. Therefore, understanding the crucial processes in simulating ENSO with a comprehensive CGCM is still fragmented. However, the ultimate goal of ENSO studies using a complicated CGCM should be to narrow the uncertainty of the ENSO simulation in terms of its amplitude, frequency, and structure by generalizing the causes of diversity in simulations. Such efforts will be made in the forthcoming Climate Model Intercomparison Project Phase 5 (CMIP5) and will contribute to reliable prediction of future changes in ENSO properties.

Acknowledgments. We are grateful to Fei-Fei Jin and Eric Guilyardi for their constructive comments. Comments by Ben Kirtman and an anonymous reviewer were also appreciated. This work was supported by the Innovative Program of Climate Change Projection for the 21st Century (“Kakushin” program) from MEXT, Japan.

APPENDIX

Mean State and Its Stability in a Simple Tropical Climate Model

To obtain insights into the dynamical relationship between the mean state change and the ENSO amplitude change in MIROC$i$, we use a simple tropical climate model (Jin 1996, hereafter J96). This model has two degrees of freedom for the ocean mixed layer temperature (or simply SST) in the eastern basin, $T_e$, and the departure of the thermocline depth from a reference depth in the western basin, $h_w$. The J96 model is regarded as a prototype to the recharge oscillator model (Jin 1997) and has the advantage that the solutions are obtained for the full state but not for the anomaly from the prescribed mean state.

The eastern basin SST is controlled by

$$\frac{dT_e}{dt} = -\varepsilon_T(T_e - T_r) - \theta(w) \frac{T_e - T_{xe}}{H_m}, \quad (A1)$$

where $\varepsilon_T^{-1}$ defines the time scale at which $T_e$ is restored to a radiative–convective equilibrium temperature ($T_r$), $T_{xe}$ is the subsurface temperature, $w$ the upwelling velocity, $H_m$ the prescribed mixed layer depth, and $\theta(x)$ the Heaviside function of $x$. The first term on the rhs of (A1) mimics the net heating due to radiative and surface heat fluxes, while the second term represents the dynamical cooling due to equatorial upwelling.

The thermocline depth in the western basin slowly adjusts to the zonally integrated Sverdrup mass transport that is proportional to the zonally integrated wind stress curl due to the basinwide dynamical adjustment at a rate $r$,

$$\frac{dh_w}{dt} = -r h_w - \frac{r b_L \tau}{2}, \quad (A2)$$

where the second term on the rhs represents the zonally integrated Sverdrup mass transport and the factor $b_L$ ($L$ is the basin length) measures the effective Bjerknes feedback mentioned in section 3a. Other parameters and diagnostic equations for $w$, $\tau$, and $T_{xe}$ are given in J96 and Watanabe (2008).

The control parameters adopted for the present purpose are $\varepsilon_T^{-1}$ and $b_L$ multiplied by a coupling coefficient $\mu$, that is, $\mu b_L$. Stationary solutions of the simple model (A1)–(A2) are calculated by an iterative solver with changing values of these two parameters, and a linear stability analysis is performed for the stationary states. The stationary solution for $T_e$ is displayed in Fig. A1a, and the growth rate (shading) and period (contours) of the linear perturbation with respect to the stationary solution is presented in Fig. A1b. The standard values of parameters used in J96 ($\varepsilon_T^{-1} = 75$ days and $\mu b_L = 12.5 \text{ m K}^{-1}$) are denoted by circles.

The stationary $T_e$, plotted on the two-dimensional parameter space, varies from a warm state ($>28^\circ C$) with smaller values of both $\varepsilon_T^{-1}$ and $\mu b_L$ to a cold state ($<20^\circ C$) with larger parameter values (Fig. A1a). The observed mean SST in the Niño-3 region lay in between and also within the range of the mean SST in the four MIROC experiments. As presented in Fig. 7, the mean SST in the cold tongue is warmest in L500 (lower dashed curves in Fig. A1a) and coldest in L575 (upper dashed curve).

The linear stability analysis shows that the stationary state is either unstable or weakly stable for most of the parameter sets (Fig. A1b). For the range of the parameters corresponding to a realistic mean SST (i.e., between two dashed curves), the perturbation has a periodicity of about 3 yr, which is reminiscent of ENSO. The mode is most unstable for large $\mu b_L$ and small $\varepsilon_T^{-1}$, arising from a low mean SST accompanying a strong zonal SST gradient.
The direction of the mean state change from L500 to L575 is not known in the parameter space, but it is assumed to be proportional to \( \epsilon T^{-1} \) (toward the cold mean state) and inversely proportional to \( \mu bL \) (toward a weak Bjerknes feedback). Along the direction shown by a red arrow, it turns out that the linear stability varies from unstable to stable regimes (Fig. A1b), indicating a change from strong to weak ENSO-like variability in the simple model. In other words, a strong ENSO can coexist with the mean state having a weak zonal SST gradient and vice versa, which is actually confirmed by integrating (A1) and (A2) over time (not shown). This result supports the arguments in section 3a and also suggests that the altering Bjerknes feedback associated with the changing zonal structure of ENSO is a necessary piece for understanding the change in ENSO amplitude in MIROC.

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