Parameterization of Tropical Instability Waves and Examination of Their Impact on ENSO Characteristics

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ABSTRACT

The impact of tropical instability waves (TIWs) on El Niño–Southern Oscillation (ENSO) characteristics is investigated by introducing a new parameterization of TIWs into an atmosphere–ocean general circulation model (AOGCM), the Model for Interdisciplinary Research on Climate (MIROC), with a medium-resolution (~1.4°) ocean model (known as MIROCmedres). Because this resolution is not sufficient to reproduce eddies at the spatial scale of TIWs, this approach isolates TIW effects from other factors that can affect ENSO characteristics. The parameterization scheme represents the effect of baroclinic eddy heat transport by TIWs. A 100-yr integration reveals a significant role of TIWs in observed ENSO asymmetry. Asymmetric heat transport associated with TIWs that are active (inactive) during La Niña (El Niño) generates a significant asymmetric feedback to ENSO and explains the observed asymmetric feature of a stronger-amplitude El Niño and weaker-amplitude La Niña. Furthermore, the parameterized eddy heat flux also affects the mean subsurface heat balance via the shallowing and steepening thermocline. This change in subsurface stratification induces a stronger thermocline feedback and a longer ENSO period.

1. Introduction

Tropical instability waves (TIWs) are observed in both the tropical Atlantic (Düeising et al. 1975) and Pacific (Legeckis 1977) Oceans as cusp-shaped frontal waves just north of the equator. In the Pacific Ocean, they develop from July to the year’s end and propagate westward at a wavelength range of 1000–2000 km and periodicity range of 20–40 days (Qiao and Weisberg 1995). The mechanism that generates TIWs is known as a mixed barotropic–baroclinic instability. Some studies have discussed the origins of TIWs by analyzing their energetics using observational data. The instability of intense latitudinal barotropic shears is associated with the shear between the Equatorial Undercurrent (EUC) and the South Equatorial Current (SEC) north of the equator (Qiao and Weisberg 1995), and between the SEC and the North Equatorial Countercurrent (NECC) (Flament et al. 1996). Baroclinic instability, where mean potential energy is converted to eddy kinetic energy, occurs at the equatorial front between 3° and 6°N, owing to the cold tongue and intertropical convergence zone (ITCZ) (Hansen and Paul 1984; Wilson and Leetmaa 1988). Numerical models have also supported these generation hypotheses (Philander 1976, 1978; Cox 1980; Semtner and Holland 1980).

A detailed understanding of TIWs is important because of their potential significance in the climatological equatorial ocean heat budget (Hansen and Paul 1984; Bryden and Brady 1989; Bauturin and Niiler 1997; Jochum and Murtugudde 2006). The equatorward heat flux of TIWs is comparable to atmospheric heat flux and offsets a considerable fraction of heat transport by mean advection. Weisberg and Weingartner (1988) also noted that TIWs reduce the shears of mean oceanic currents.

Recently, several studies have focused on the relationship between TIWs and El Niño–Southern Oscillation (ENSO). Yu and Liu (2003), using an atmosphere and ocean general circulation model (AOGCM), found a linear relationship between ENSO variability and interannual variation in TIW intensity. They showed that TIW intensity increases during La Niña years and is reduced during El Niño years with a linear relationship between the year-to-year variations of TIW activity and interannual Niño-3 (5°S–5°N, 150°–90°W) sea surface temperature (SST) anomalies. Their correlation analysis suggested that TIWs are modulated by ENSO through the SST gradient north of the equator, which is related to
the mechanism generating baroclinic instability. These results raise questions as to how the subsequent changes in TIW-induced heat transports affect the thermodynamic structure of the ITCZ–cold tongue complex and feedback to the ENSO cycle.

Vialard et al. (2001) showed the importance of the TIW contribution to the surface-layer heat budget at an ENSO time scale. They closely examined the mixed layer heat balance for the 1997/98 El Niño by forcing their tropical Pacific Ocean GCM with a combination of weekly European Remote Sensing (ERS-1–2) satellite data and Tropical Ocean Global Atmosphere-Tropical Atmosphere Ocean Project (TOGA-TAO) wind stress data. They concluded that the interannual anomaly of the eddy contribution appears as a significant negative feedback to the growth of El Niño events because an interannual component of TIW-induced heat transport always works as a damping process in ENSO growth. In other words, during an El Niño, TIW-induced equatorward heat transport is weaker than usual and slows the growth of El Niño, as TIWs are much weaker because of diminished shear between the SEC and NECC and a diminished meridional temperature gradient.

An (2008) provided a new perspective on the feedback of TIWs to ENSO by suggesting that TIWs could play an important role in producing the asymmetry between warm and cold ENSO events. As shown by Vialard et al. (2001), TIW variability acts as a negative feedback to ENSO. However, this feedback is stronger during La Niña than during El Niño because the temperature gradient to the north of the cold tongue is greater, and TIWs are more active during La Niña. Such a negative/ asymmetric feedback results in El Niño–La Niña asymmetry, with larger amplitude during El Niño than during La Niña. An (2008) identified this nonlinear process by incorporating a simple statistical parameterization of the horizontal heat flux convergence (HFC) due to TIWs into a simple scalar ENSO model consisting of two equations representing the tendencies of western thermocline depth changes and eastern SST anomalies in the equatorial Pacific. The numerical results of Sakamoto et al. (2012) also support this hypothesis. By comparing the performances of three AOGCMs with different resolutions for the atmosphere and ocean, they found that a high-resolution ocean model that can reproduce TIWs showed stronger asymmetry of Niño-3 indices than did a low-resolution model.

Observed El Niño events are often stronger than La Niña events, and particularly strong El Niño events have occurred during recent decades. In addition to TIWs in the equatorial eastern Pacific, several nonlinear processes in the tropical air–sea coupled system may cause the asymmetric behavior of ENSO, such as nonlinear dynamical heating (NDH) of ENSO (An and Jin 2004), the vertical mixing process in the ocean mixed layer, the asymmetric response of the atmosphere to SST anomalies (Kang and Kug 2002), and thermodynamic control of deep convection (Hoerling et al. 1997). Their relative contributions remain unresolved. The ENSO model presented by An (2008) is too simple to confirm the TIW contribution in the presence of other potential processes causing ENSO asymmetry.

This study investigates how TIW-induced heat transports feed back to the ENSO cycle, including the other nonlinear effects on El Niño–La Niña asymmetry. An OCGM only is insufficient because ENSO and its asymmetry develop through global atmosphere and ocean interactions. Long-term integration of an AOGCM is necessary to discuss the impact of TIWs on ENSO characteristics. However, a high-resolution AOGCM is not necessarily appropriate to investigate the significance of TIW effects on ENSO characteristics because the reproduced ENSO would be affected by both TIWs and other climate components, making identification of causality difficult. In this study, we introduce a new TIW parameterization into a medium-resolution AOGCM, which alone cannot resolve TIWs because of not only the direct disadvantage of their coarse resolution but also the underestimation of zonal current shears due to numerical damping. This approach enables us to detect the pure impact of TIWs on the ENSO characteristics. Section 2 describes the model and parameterization method in detail. In section 3, the impact of TIWs on oceanic mean fields is presented. The impact of TIWs on ENSO, which is the main focus of this study, is investigated in section 4 using the parameterized AOGCM. Section 5 provides the discussion and summary.

2. Methodology

2.1. Model

The coupled AOGCM used in this study is the Model for Interdisciplinary Research on Climate (MIROC), version 5.0i, which is based on MIROC3 (K-1 Model Developers 2004). The model includes atmosphere, land, river, sea ice, and ocean. The ocean and atmosphere are coupled at 3-h intervals without applying any flux adjustment.

The atmospheric part of the model is based on the primitive equations for a sphere, using a spectral-transform method for horizontal discretization. Vertical sigma coordinates are used. The horizontal resolution is horizontal triangular spectral truncation at total wavenumber 42 (T42) with 40 vertical layers. The physical parameterizations
incorporate a higher-order 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]. The convection scheme of Chikira and Sugiyama (2010) is newly developed to control the simulated ENSO amplitude by changing the efficiency of the entrainment rate in cumuli (Watanabe et al. 2011).

The ocean part of the model is the Center for Climate System Research (CCSR; now Atmosphere and Ocean Research Institute, AORI) ocean component model (COCO; Hasumi 2006). The model is based on the primitive equations with Boussinesq approximation. The horizontal resolution is 1.4° longitude and 0.9° latitude (0.5° near the equator). There are 44 vertical levels spanning 5 m in the top level and 250 m in the bottom level. The turbulence closure scheme of Noh and Kim (1999) is used to calculate vertical viscosity and diffusion coefficients. The model also incorporates the uniformly third-order polynomial interpolation algorithm (UTOPIA; Leonard et al. 1994) for tracer advection and formly third-order polynomial interpolation algorithm.

The eddy-induced transport of tracers is parameterized by the isopycnal layer thickness diffusion (Cox 1987). Eddy-induced trans-

sion coefficients. The model also incorporates the uni-

formly third-order polynomial interpolation algorithm (UTOPIA; Leonard et al. 1994) for tracer advection and the isopycnal diffusion (Cox 1987). Eddy-induced trans-
port of tracers is parameterized by the isopycnal layer thickness diffusion of Gent and McWilliams (1990).

In this study, the model is integrated for 85 yr under preindustrial climate forcing of solar radiation, volcanism, greenhouse gases, ozone, aerosol, and land-use change, which were set to values corresponding to the year 1850 (CTL run). As a comparison run, the same integration is conducted, except that the baroclinic eddy parameterization is introduced (TIW run). Both experiments are started from the same initial conditions, which are randomly selected from the several-thousand-year control integration. The first 8 yr are discarded from the analysis. In the following sections, monthly outputs are analyzed and compared between the CTL and TIW runs.

b. Tropical instability wave parameterization

To consider the impact of the eddy heat budget induced by TIWs, mesoscale eddy parameterization of isopycnal-layer thickness diffusion is adopted into the medium-resolution MIROC. This parameterization was originally developed for baroclinic fronts in mid- or high-latitude \( f \) planes, such as the Ekman convergence zone in the subtropical North Atlantic and North Pacific and the Antarctic Circumpolar Current (Bryan et al. 1999; Gordon et al. 2000). This is the first time that this parameterization has been used for TIWs.

The effect of mesoscale eddies on tracer transport, which reduces the slope of the isopycnal surfaces, is often parameterized by eddy-induced transport velocity associated with isopycnal layer thickness diffusion (Gent and McWilliams 1990; Gent et al. 1995), which is incorporated into the MIROC ocean model. The original scheme employs a constant coefficient for the isopycnal layer thickness diffusion, while actual mesoscale eddy activities are inhomogeneous, temporally and spatially. Visbeck et al. (1997) proposed a modified thickness diffusion scheme, with \( K_G \) being a variable coefficient dependent on local baroclinicity. The definition of \( K_G \) includes the eddy length scale and eddy growth rate as follows:

\[
K_G = \alpha \frac{L^2}{T},
\]

where \( \alpha \) is a constant of proportionality, \( L \) is the length scale of eddy transfer, and \( T \) is the time scale of eddy growth. The constant parameter \( \alpha \) depends on each model. Here, we set \( \alpha \) to 0.075 after several sensitivity tests.

The eddy growth rate, \( 1/T \), is given by the Eady growth rate (Eady 1949), based on a geostrophic mean flow as follows:

\[
\frac{1}{T} = \frac{f}{\sqrt{R_i}},
\]

where \( f \) is the Coriolis parameter and \( R_i \) is the Richardson number of a large-scale field. In thermal wind balance, the Richardson number of the large-scale flow is given by

\[
R_i = \frac{N^2}{\left(\frac{\partial u}{\partial z}\right)^2} = \frac{f^2N^2}{M^4},
\]

where \( \partial u/\partial z \) is the vertical shear of mean horizontal velocity; \( N^2 \) and \( M^2 \) are measures of vertical and horizontal stratification as \( N^2 = \partial b/\partial z \) and \( M^2 = |\partial b/\partial x| \), respectively; and \( b \) is the buoyancy. As a result, the growth rate of an Eady wave can be written as

\[
\frac{1}{T} = \frac{f}{\sqrt{R_i}} = \frac{M^2}{N} = s_p N,
\]

where \( s_p = M^2/N^2 \) is the slope of an isopycnal surface. The growth rate given by Eq. (4) has a large value when horizontal stratification and weak vertical stratification exist, that is, where isopycnal surfaces are steeply sloping, such as in the northern front of the cold tongue where TIWs are active.

The length scale of the eddy transfer, \( L \), is also an important factor. Generally, the internal Rossby radius of deformation, \( \lambda_R \), is adopted for \( L \) in mid- or high-latitude \( f \) planes. However, \( \lambda_R \) overestimates the length
scale of eddy transport in the equatorial $\beta$ plane. Therefore, in this study, the Rhines scale proposed by Held and Larichev (1996) is adopted, where the scaling argument on an $f$ plane is extended to a $\beta$ plane, and the barotropic inverse energy cascade can be considered. The Rhines scale is given as

$$L = \sqrt{\frac{U}{\beta}}$$

(5)

where $U$ is the horizontal velocity of mean flow and $\beta$ is a planetary beta. The order of the Rhines scale in an active TIW region is 1000 km, which is reasonable for a TIW scale. The parameterized thickness diffusion coefficient is applied in tropical regions between 15°S and 15°N down to 200-m depth and linearly decays outside these latitudes toward the original constant value.

c. Performance verification of mesoscale eddy parameterization

Figure 1 shows the distribution of the isopycnal layer thickness diffusion coefficient, $K_G$, calculated by the new parameterization. Large values are distributed over the region of TIWs, whereas the coefficient is set to a constant value in the original MIROC CTL run. To verify whether these values are reasonable, we estimate a meridional eddy HFC from $K_G$, and compared it with HFC from a high-resolution ocean model [Fig. 2; details of the TIW simulation with the high-resolution ocean GCM are described by Imada et al. (2012)]. In the medium-resolution MIROC, the HFC by eddies is calculated depending on $K_G$ and isopycnal diffusion, $K_I$, as follows (see the appendix):

$$
\begin{pmatrix}
K_I & 0 & (K_I - K_G)s_x & \left(\frac{\partial T}{\partial x}\right)
0 & K_I & (K_I - K_G)s_y & \left(\frac{\partial T}{\partial y}\right)
(K_I + K_G)s_x & (K_I + K_G)s_y & K_I(s_x^2 + s_y^2)s_z & \left(\frac{\partial T}{\partial z}\right)
\end{pmatrix}
$$

(6)

where $s_x = -\left(\frac{\partial \rho}{\partial x}\right)/(\partial \rho/\partial z)$ and $s_y = -\left(\partial \rho/\partial y\right)/(\partial \rho/\partial z)$. The high-resolution COCO model (resolution: 0.28° longitude, 0.18° latitude) directly estimated $-\left[\partial(v^\prime T^\prime)/\partial y\right]$. The horizontal distribution and values simulated by both models are comparable, with HFC shown as positive around the equator and negative between 2° and 5°N. However, large values are located too far east in the TIW run, as the mean thermocline is too tight, due to the lack of resolution.1 In the vertical section, large positive and negative values are found across the steep slope of the mean thermocline, and those values are on the same order in both models. Therefore, our new scheme for TIWs operates appropriately.

3. Impact on mean climatology

First, we examined the impact of the mesoscale eddy parameterization introduced in this study on oceanic mean fields. It is known that a tropical SST cold bias due

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1 The absence of the Galapagos Islands in a medium-resolution MIROC is one of the major reasons for the too-tight thermocline in the equatorial eastern Pacific. Karnauskas et al. (2007) showed that the existence of the Galapagos Islands acts as the obstruction of the EUC that is related to the subduction on the equatorial farther-eastern Pacific. As a result, the tilt of the thermocline relaxes and SST cold biases are reduced around the eastern edge of the tropical Pacific Ocean in the presence of Galapagos Islands. In the coarser-resolution models without the geography of the Galapagos Islands, because this subduction cannot appear, the thermocline layers are concentrated at the surface.
to the bias of the westward shift of the cold tongue, which is a problem in many AOGCMs, is reduced in the higher-resolution oceanic model, as the equatorward heat transport associated with TIWs is realistically simulated (Roberts et al. 2009; Shaffrey et al. 2009). This tendency of diminished SST biases is also represented in the high-resolution MIROC version, compared with the low-resolution version (Sakamoto et al. 2012).

The changes in mean SST resulting from the introduction of the TIW parameterization are shown in Fig. 3a. The original MIROC (CTL run) has cold SST biases along the equatorial Pacific Ocean. However, the biases are reduced over a wide area in the TIW run, which is consistent with the conclusions of Roberts et al. (2009). Figure 3b shows the latitude–depth section of the ocean temperature changes in the eastern Pacific associated with the new parameterization, along with the contour lines of mean temperature simulated by the CTL run. A strong meridional gradient of mean temperature exists on both sides of the equator from the surface to 50-m depth. Near the surface, a warming tendency is more dominant on the equator side of this temperature front than outside it, as equatorward heat transport is induced by the effects of parameterized mesoscale eddies. On the other hand, a cooling tendency exists on the equator side below 100-m depth. This is because the inverse mean temperature front, with warmer temperatures on the equator side and cooler temperatures on the northern side, is evident around this depth. Such an inverse mean temperature front is also identified in the observed stratification, while MIROC tends to overestimate the meridional temperature gradient. Parameterized heat transport works to weaken this inverse front with cooling on the equator side and warming on the off-equator side. Therefore, the introduced TIW scheme plays a role in reducing the biases in both the surface and subsurface regions.

Finally, the resultant changes in stratification show the cooling trend along the original (CTL run’s) thermocline and the warming trend near the surface. As a result, the vertical temperature gradient is intensified above the original thermocline; that is, the thermocline shifts upward. Figure 3c shows the differences in the standard deviation (SD) of interannual ocean temperature anomalies averaged between 2°S and 2°N between the TIW and CTL runs, which correspond to changes in interannual temperature variability resulting from the inclusion of the baroclinic eddy effect. The mean temperature of the CTL run is also indicated by broken lines. Variability is enhanced just above the original (CTL run’s) equatorial thermocline, indicating the thermocline shallows because temperature variability is larger where the steeper temperature gradient exists. Variability also increases near the surface, but this increment is not as evident in the eastern Pacific as around the thermocline, because isopycnal layer thickness diffusion, represented by $K_G$, has a local maximum near the surface (Fig. 1) and damps ENSO amplitude. Because the simulated equatorial thermocline is too broad and deep in the MIROC simulations (Imada and Kimoto 2006), the TIW scheme
introduced in this study also has the potential to reduce these biases.

4. Impact of tropical instability waves on ENSO

a. Impact on ENSO skewness

Figure 4 shows the time series of SST anomalies in the Niño-3 region (5°S–5°N, 150°E–90°W) calculated from the CTL and TIW runs. In the TIW run, large El Niño events are noticeable, whereas La Niña events are not as large and terminate slowly. This tendency is not obvious in the CTL run. Figure 5 shows probability density distribution histograms. The CTL run index is distributed almost symmetrically, whereas the TIW run distribution deviates from a normal distribution, with its index skewed to warm values.

To quantify the deviations from normality, we examine skewness, $\sqrt{b}$, using the following equation:

$$\sqrt{b} = \frac{m_3}{m_2^{3/2}},$$

where $m_k = \frac{(x - \bar{x})^k}{n}$ is the $k$th moment about the mean (Mardia 1980; Burgers and Stephenson 1999). Skewness toward high values indicates that high extremes are
more probable than low extremes, and \( \sqrt{b} = 0 \) for a large enough sample from a normal distribution. Niño-3 SST anomalies from the 1950–97 ocean reanalysis dataset of Ishii and Kimoto (2009) yield \( \sqrt{b} = 0.89 \), which indicates that large El Niño events are more frequent than large La Niña events. The Niño-3 index from the TIW run has a large positive skewness of \( \sqrt{b} = 0.95 \). On the other hand, the Niño-3 index from the CTL run shows a more normal distribution of \( \sqrt{b} = 0.34 \). This comparison also confirms the role of baroclinic eddy heat flux in El Niño–La Niña asymmetry.

Figure 6a presents the spatial distribution of skewness for observed SST anomalies in the Pacific Ocean. Skewness decreases from east to west across the tropical Pacific. Large SST skewness occurs in the eastern equatorial Pacific, where the thermocline is close to the surface and, therefore, resembles the distribution of the ENSO anomalies. Figures 6b,c show the skewness from the CTL and TIW runs, respectively. In the CTL run, the asymmetry of the interannual SST variability is overestimated in the central tropical Pacific and underestimated in the region where ENSO occurs. One possible reason for the asymmetry in the central Pacific appears to be the asymmetric physical process of cumulus convection. As pointed out by Hoerling et al. (1997), the center of convection anomalies during La Niña is shifted to the west, compared to that during El Niño, as a negative SST anomaly in the cold tongue during La Niña has no further effect on convection. This asymmetric cumulus convection induces the asymmetry of SST variability through the wind–SST feedback. Therefore, the asymmetric response of convection on SST is one possible reason for the westward shift of the SST skewness in the CTL run shown in Fig. 6b, because the version of MIROC used in this study tends to overestimate the wind–SST feedback.

On the other hand, the distribution in the TIW run differs from that in the CTL run, as skewness increases significantly in the eastern tropical Pacific owing to the effect of mesoscale eddy heat flux associated with TIWs, indicating that the TIW-induced asymmetric SST fluctuation contributes significantly to the improvement in the simulated ENSO asymmetry. The impact of the TIW parameterization is dominant in the eastern Pacific Ocean where the thermocline surfaces because this parameterization calculates large isopycnal layer thickness diffusion in the region where an intense horizontal density gradient exists, such as along the off-equatorial thermocline.
To investigate the feedback mechanism of TIWs on ENSO, we calculate meridional HFC due to TIWs in the eastern tropical Pacific upper 50-m depth from the parameterized coefficient $K_G$ [see Eq. (4)]. To compute the interannual effect of TIWs, the seasonal cycle of HFC, which modifies the seasonal SST budget, is removed from the original HFC. Figure 7 shows a scatterplot of Niño-3 SST anomalies versus HFC anomalies. A linear regression between the Niño-3 index and HFC anomalies is calculated separately for the positive and negative Niño-3 indices. For comparison, this calculation is also conducted for all the data. In the TIW run, the slope of the regression line associated with the negative Niño-3 index (La Niña years) is steeper than that associated with the positive Niño-3 index (El Niño years). That is, the warming tendency by HFC anomalies during La Niña is greater than the cooling tendency by HFC anomalies during El Niño. This tendency is the same as that found by An (2008; see his Fig. 3), who used TOGA-TAO data and the Simple Ocean Data Assimilation (SODA). On the other hand, the CTL run shows an unrealistic relationship, as the slope is rather steeper during El Niño years than during La Niña years. This also supports the effectiveness of the introduction of the baroclinic eddy parameterization for TIWs.

b. Impact on an ENSO mode

Previous studies have described two types of modes that can give rise to ENSO (Neelin et al. 1998; Fedorov and Philander 2000; Burgers and van Oldenborgh 2003; Wang and Picaut 2004): an SST mode resulting from local SST–wind interactions in the central-eastern Pacific and a thermocline mode resulting from remote winds–thermocline feedbacks in the western Pacific. The SST mode is associated with east-to-west propagation of SST anomalies and low-amplitude ENSO events at a 2–3-yr frequency, whereas the thermocline mode has west-to-east propagation of subsurface temperature anomalies and large events at a 4–5-yr frequency (Guilyardi 2006). The SST mode is controlled by the anomalous zonal advection of mean temperature gradients, while the thermocline mode is primarily associated with the vertical advection of temperature anomalies by the mean upwelling.

Those components are included in a SST anomaly budget equation (Kang et al. 2001; Imada and Kimoto 2006). Figure 8 shows lag-regressing anomalous zonal advection and mean upwelling advection terms averaged over the central-to-eastern tropical Pacific ($5^\circ S$–$5^\circ N$, $170^\circ W$–$90^\circ W$) onto SST anomalies. The zero lag corresponds to the mature phase of ENSO. For reference, a lag regression for the SST tendency is also plotted. In the CTL run, the anomalous zonal advection term leads ENSO progression and is in phase with the SST tendency, while the mean upwelling advection term is smaller and lags behind the SST tendency. These results indicate that a SST mode is more in effect in the CTL run than a thermocline mode. On the other hand, in the TIW run, the mean vertical advection term is in phase with the SST tendency. Though the anomalous zonal advection still shows large value, it peaks after the maximum phase of the SST tendency. Therefore, the thermocline mode is enhanced and contributes to ENSO development along with the SST mode in the parameterization run. Because the observed ENSO also presents a mixed mode, a more realistic ENSO mode comes into the TIW run.

The enhancement of the thermocline mode efficiency is explained as a result of changes in mean stratification due to the existence of TIWs discussed in section 3. Figure 3c shows that temperature variability is enhanced just above the equatorial thermocline while the increment is not as evident near the surface as around the thermocline. As a result, temperature variability induced by a thermocline mode is likely to be relatively enhanced in the TIW experiment.
c. Impact on ENSO frequency

Figure 9 shows the power spectra of Niño-3 indices from the CTL and TIW runs. The spectra clearly show that the ENSO period simulated by the TIW run is longer than that of the CTL run. Three possible reasons may explain why baroclinic-eddy parameterization affects ENSO frequency. One is due to the strengthened...
thermocline mode in the TIW run, as described in the previous section, because the thermocline mode has a relatively longer ENSO period.

The second possible reason involves a depth change in equatorial Rossby wave propagation. Figure 3d shows the difference in SD of temperature variability, as in Fig. 3c, but averaged between 5° and 6°N. After introducing the TIW scheme, temperature variability is largely reduced around the thermocline and increased just above the depth. These results indicate that the depth of the thermocline also shallows in these latitudes. The depth of the layer influences the phase speed of the equatorial waves. The Rossby wave dispersion relationship in the shallow-water approximation derives a positive relationship between Rossby wave phase speed and layer thickness. Therefore, the shallower the depth of the thermocline, the slower its phase speed becomes. Figure 10 is the longitude–time diagram of the lag-regressing temperature anomaly at off-equator averaged from 150- to 300-m depth onto the Niño-3 SST anomalies. Just after the mature phase of El Niño (at zero lag), negative anomalies propagate westward, which corresponds to the equatorial Rossby wave. In the TIW run, this propagation takes more time with a slower phase speed than in the CTL run. As a result, the slow Rossby wave propagation can give a longer ENSO period in the TIW run.

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**Fig. 9.** Power spectrum of Niño-3 indices from the CTL (solid line) and TIW (dashed line) runs. Dots show red noise spectrum levels.

**Fig. 10.** Longitude–time diagram of lag-regressing temperature anomaly (K) averaged from 150- to 300-m depth and between 5° and 6°N onto the Niño-3 index. The zero lag corresponds to the mature phase of ENSO. Positive values are shaded: (left) CTL and (right) TIW runs.
The spatial pattern of the wind stress anomalies can be the other possible reason. Some previous studies have showed that the increase in the ENSO period is due to the eastward shift or an expanding meridional scale of zonal wind stress anomalies with respect to the SST anomalies (An and Wang 2000; Capotondi et al. 2006). Figures 11a,b show the pattern of zonal wind stress responses to ENSO from two simulations. At the mature phase of El Niño, the westerly wind stress anomaly covers the equatorial and off-equatorial Pacific. The peak of the westerly wind stress anomaly is located around 170°W in the CTL run (Fig. 11a) and around 150°W in the TIW run (Fig. 11b). This difference is because the model bias of an exaggerated westward extent of the cold tongue is diminished by introducing the TIW parameterization (Fig. 3a). An associated ENSO position bias that SST anomalies tend to shift to the western and central Pacific, which is a known common bias among many AOGCMs (e.g., Latif et al. 2001), is also diminished in the TIW run (Fig. 11c: a regression of SST anomalies on the Niño-3 index). An and Wang (2000) indicated that the longitudinal position of the zonal wind stress anomalies affects the ENSO period by influencing the phase relationship of the zonal advection with SST anomalies. According to them, when wind stress anomalies are displaced farther east, the anomalous zonal advection term lags behind the SST tendency (Fig. 8) and contributes to ENSO amplification rather than its phase transition, which is consistent with the hypothesis of An and Wang (2000). Therefore, the more eastern position of the wind stress anomalies in the TIW run can be one of the reasons of a longer ENSO period.

5. Discussion and summary

This study has explored the impact of TIWs on ENSO by introducing a new parameterization of a TIW-induced heat flux. An (2008) suggested that asymmetric thermal heating associated with TIWs induces a negative/asymmetric feedback to ENSO and can explain the El Niño–La Niña asymmetry of larger-amplitude El Niño and smaller-amplitude La Niña. However, previous studies have also proposed several nonlinear processes in the tropical air-sea coupled system that may cause the asymmetric behavior of ENSO. Which processes are more important in driving the asymmetry of ENSO has yet to be addressed.

The use of a coupled AOGCM containing the TIW parameterization enabled us to estimate the significance of TIWs among other possible processes causing ENSO asymmetry. After incorporating the TIW parameterization, our model results showed an increase in ENSO skewness in the eastern Pacific and supported the

![Fig. 11. Zonal wind stress anomalies regressing on the Niño-3 index (N m$^{-2}$) from the (a) CTL and (b) TIW runs. Positive values are shaded. (c) Regression of SST anomalies on the Niño-3 index (K) averaged between 2°S and 2°N from the CTL (broken line) and TIW (solid line) runs.](image-url)
significant impact of TIW-induced heat transport on the asymmetry of ENSO, in addition to other effective factors.

This numerical study also clarified the impact of baroclinic eddies on ENSO modes and frequency. TIW-induced heat convergence tends to modify the stratification around the tropical thermocline, with the thermocline becoming shallower and steeper. As a result, the contribution of a thermocline mode to ENSO development is enhanced. Furthermore, the ENSO cycle is also lengthened by considering the eddy heat transport. We suggest a few possible reasons: enhancement of a thermocline mode, which propagates at longer frequency than that of a SST mode; the shallower off-equatorial thermocline, which slows the propagating equatorial Rossby wave; and eastward shift of wind anomalies due to the improved spatial pattern of ENSO variability, which also lengthens the ENSO cycle.

Here, we should reconsider the reason of enhanced ENSO asymmetry in the parameterization run, because TIW-induced heat convergence affected not only ENSO asymmetry but also the mean stratification and the other ENSO characteristics such as its spatial pattern. Modified ENSO characteristics may indirectly impact on ENSO asymmetry because the NDH of ENSO is known to be one of the major contributors of ENSO asymmetry (An and Jin 2004). To evaluate the relative contribution of HFC by TIWs against that of NDH associated with ENSO, the NDH was also estimated for the CTL and TIW runs (Fig. 12) in the same unit system as in Figs. 2a,b. Figure 12 shows the increase in NDH in the eastern equatorial Pacific after introducing our TIW scheme. However, the increment is slight compared to the parameterized HFC [Fig. 2b; note that estimated HFC for the CTL run by Eq. (6) is negligible compared to Fig. 2b because the constant value of $K_G$ in the CTL run is much smaller than the parameterized $K_G$ in the TIW run]. This also supports a robust impact of TIW-induced heat transport on ENSO asymmetry.

All of our results indicate that TIW effects contribute significantly to ENSO characteristics. At present, most AOGCMs used for seasonal-to-interannual prediction do not reproduce TIWs because of a lack of oceanic resolution. Ham and Kang (2011) examined the impact of TIWs on the initial conditions on seasonal forecasts by AOGCM, and confirmed improvement in the prediction skill for the Niño-3 index. Hence, the TIW parameterization developed in this study has the potential to improve our skill in forecasting ENSO events.

Observational data have shown that the asymmetry of ENSO has increased concurrent with the 1970s climate shift (Wu and Hsieh 2003; An 2004; An and Jin 2004; An et al. 2005). An and Jin (2004) reported that the interdecadal changes in ENSO nonlinearity were related to SST climate-state change and subsurface temperature. Many studies have shown that the SST pattern change in the 1970s has produced local maximum values in the off-equatorial eastern Pacific in both the Northern and Southern Hemispheres (e.g., Imada and Kimoto 2009); that is, the increasing meridional SST gradient at the north of the cold tongue is favorable for the activation of TIWs. Considering all these factors, the new view of TIW activity may provide a clue for investigating the relationship between the mean climate shift and the modulation of ENSO asymmetry. The medium-resolution MIROC used in this study can reasonably reproduce interdecadal ENSO-like variations and the related decadal modulation of ENSO (Imada and Kimoto 2009). Thus, this version of MIROC with TIW parameterization incorporated may aid in investigating new mechanisms of interdecadal variability impacts on ENSO characteristics via the effect of TIWs. However, long-term integration beyond the scope of this study is required to further discuss this topic.

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APPENDIX

Equation for Eddy Heat Flux Convergence

To remap diffusive tracer fluxes along isopycnal-diapycnal surfaces to fluxes in the x, y, and z directions, the formulation of coordinate system rotation is required. The approximation method by Cox (1987) is adopted in this study. The slope of the isopycnal surface in the x and y directions is given by

\[ s_x = -\frac{\partial x}{\partial x} \frac{\partial P}{\partial x} \quad s_y = -\frac{\partial x}{\partial y} \frac{\partial P}{\partial z} \]  

(A1)

respectively. The rotation matrix along the isopycnal surface becomes

\[
\begin{pmatrix}
1 & 0 & s_x \\
0 & 1 & s_y \\
s_x & s_y & s_x^2 + s_y^2
\end{pmatrix}
\]  

(A2)

and in the diapycnal direction, the rotation matrix becomes

\[
\begin{pmatrix}
0 & 0 & -s_x \\
0 & 0 & -s_y \\
s_x & s_y & 0
\end{pmatrix}
\]  

(A3)

Let \( K_I \) be a coefficient for the isopycnal diffusion. Diffusive fluxes of tracer \( T \) in the x, y, and z directions are represented by

\[
\begin{pmatrix}
K_I \frac{\partial T}{\partial x} + s_x \frac{\partial T}{\partial z} \\
K_I \frac{\partial T}{\partial y} + s_y \frac{\partial T}{\partial z} \\
K_I \left[ s_x \frac{\partial T}{\partial x} + s_y \frac{\partial T}{\partial y} + (s_x^2 + s_y^2) \frac{\partial T}{\partial z} \right]
\end{pmatrix}
\]  

(A4)

Let \( K_G \) be a coefficient for isopycnal layer thickness diffusion. Diffusive fluxes of tracer \( T \) in the x, y, and z directions are represented by

\[
\begin{pmatrix}
-K_G s_x \frac{\partial T}{\partial z} \\
-K_G s_y \frac{\partial T}{\partial z} \\
K_I \left[ s_x \frac{\partial T}{\partial x} + s_y \frac{\partial T}{\partial y} \right]
\end{pmatrix}
\]  

(A5)

Therefore, the combined coefficient tensor of the isopycnal diffusion and layer thickness diffusion becomes

\[
\begin{pmatrix}
K_I & 0 & (K_I - K_G)s_x \\
0 & K_I & (K_I - K_G)s_y \\
(K_I + K_G)s_x & (K_I + K_G)s_y & K_I(s_x^2 + s_y^2)
\end{pmatrix}
\]  

\[
\begin{pmatrix}
\frac{\partial T}{\partial x} \\
\frac{\partial T}{\partial y} \\
\frac{\partial T}{\partial z}
\end{pmatrix}
\]  

(A6)

REFERENCES

——, 2008: Interannual variations of the tropical ocean instability wave and ENSO. J. Climate, 21, 3680–3686.