Mechanisms of Future Changes in Equatorial Upwelling: CMIP5 Intermodel Analysis

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ABSTRACT

The future change in equatorial upwelling between 1971–2000 and 2071–2100 is investigated using data from 24 coupled climate models. The multimodel ensemble (MME) mean exhibits substantial equatorial upwelling decrease in the eastern Pacific and weaker decrease in the western Atlantic Ocean. The MME mean of upwelling change and intermodel variation of that are decomposed into distinct isopycnal and diapycnal components. In the Pacific, the diapycnal upwelling decreases near the surface, associated with a weakened Ekman pumping. The isopycnal upwelling decreases at depths of 75–200 m around the core of the Equatorial Undercurrent (EUC) due to flattening of the density layer in which it flows. Both the weakened Ekman pumping and the EUC flattening are induced by the locally weakened trade wind over the eastern Pacific basin. In the equatorial Atlantic, both the change in MME mean and the intermodel variation of upwellings are significantly related to the weakened trade wind and enhanced stratification, although these drivers are not independent. The results for the Pacific Ocean imply that future reduction in upwelling may have impacts at different depths by different mechanisms. In particular, the rapid warming of sea surface temperature in the eastern Pacific basin may be mainly caused by the near-surface diapycnal upwelling reduction rather than isopycnal upwelling reduction associated EUC flattening, which is important at deeper levels.

1. Introduction

Equatorial upwelling plays important roles in physical climate and marine ecosystems. The equatorial upwelling induced by easterly winds creates cold sea surface temperatures in the eastern equatorial Pacific and Atlantic Oceans where the thermocline is shallow, and sharpens the zonal contrast of sea surface temperature (SST). This SST gradient maintains a zonal sea level pressure gradient and consequent equatorial easterly wind, forming the lower branch of the Walker circulation (Bjerknes 1969). Variability in equatorial upwelling is a key aspect of the

Bjerknes feedback, which plays an essential role in El Niño–Southern Oscillation (ENSO) (e.g., Bjerknes 1969; Zebiak and Cane 1987; McPhaden 1999; Zhang et al. 2008; Fang and Wu 2008). On interannual to decadal time scales, changes in equatorial upwelling also modulate the high surface biological productivity and air–sea CO₂ flux (Chavez et al. 1999), as well as the subsurface oxygen minimum zones (Deutsch et al. 2014). As such, future changes in physical climate and marine ecosystems due to global warming may be significantly influenced by the equatorial upwelling.

Change in equatorial upwelling under global warming has mainly been investigated by analyzing the outputs of climate models, in particular those in the World Climate Research Programme (WCRP) multimodel dataset from phase 3 of the Coupled Model Intercomparison Project (CMIP3) (Meehl et al. 2007). Vecchi and Soden (2007) analyzed outputs of CMIP3 models for the Special Report

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on Emissions Scenarios (SRES) A1B of the Intergovernmental Panel on Climate Change (IPCC). They found that upwelling of multimodel ensemble (MME) mean in the eastern equatorial Pacific declines until the year 2100. By analyzing outputs of doubled CO₂ emission experiments from CMIP3 models, DiNezio et al. (2009) showed a reduction of MME mean vertical velocity at a depth of 50 m in the equatorial Pacific. In contrast, Seo and Xie (2011) found that upwelling will intensify in the eastern equatorial Atlantic until 2050 under the SRES-A1B scenario by conducting dynamical downscaling for a CMIP3 model.

Two major mechanisms for the future change in equatorial upwelling have been proposed: change in Ekman pumping near the surface and change in the Equatorial Undercurrent (EUC) at depth. For the former, DiNezio et al. (2009) reported the weakening of MME mean Ekman divergence in the equatorial Pacific and an accompanying reduction in vertical velocity at a depth of 50 m. For the latter, Vecchi and Soden (2007) suggested that a flattening in the east–west thermocline slope induces a reduction in upwelling in the equatorial Pacific; the flatter thermocline's zonal slope is directly related to the reduction in upwelling, as the eastward-flowing EUC on eastward-shallowing pycnoclines gives rise to upwelling. The EUC flattening is also accompanied by a basinwide shoaling of the EUC, as reported by Luo and Rothstein (2011) and Sen Gupta et al. (2012) based on CMIP3 analyses. In addition to these two mechanisms, a nonlinear response of the equatorial zonal currents to the intensified cross-equatorial southerlies was suggested to enhance upwelling in the cold tongue of the equatorial Atlantic (Seo and Xie 2011).

Future changes in Ekman pumping and the EUC, which are important aspects of changes in upwelling, are presumably caused by weakening of the trade wind. Most climate models in the CMIP3 project the future slowdown of the Walker circulation under global warming (Vecchi and Soden 2007; Sen Gupta et al. 2012). The relation among the trade wind, upwelling, and zonal thermocline gradient has been established for El Niño studies (e.g., Philander 1981; McPhaden 1993; Jin 1997; Kirtman 1997; McPhaden 1999; Wittenberg 2002; Zhang et al. 2008; Liu et al. 2017). Moreover, the tendency toward a weakening of the equatorial easterlies associated with the slowdown of the Walker circulation over the last 60 years has already been observed, although the intensity of Walker circulation exhibits a substantial decadal variability and whether the observed slowdown in the Walker circulation is due to natural variability or anthropogenic global warming remains controversial (Vecchi et al. 2006; Zhang et al. 2008; Tokinaga et al. 2012a,b). These results underline the importance of understanding how equatorial upwelling responds to the change in trade wind in a warmer climate.

Although previous studies have shown that future trade wind weakening will cause a reduction in equatorial upwelling, how the aforementioned two mechanisms contribute to the weakening of upwelling at different depths and different locations has not been systematically investigated. The purpose of this study is to obtain a better understanding of the mechanisms of future changes in equatorial upwelling in the Pacific and Atlantic Oceans, where substantial reductions in upwelling are expected (Vecchi and Soden 2007). For this purpose, we systematically analyze output from 24 CMIP5 models until 2100 under representative concentration pathway (RCP) 8.5 by employing two analyses. One is to analyze the relation between MME mean and intermodel variation of future changes; this analysis can tell us how much of the change in MME mean can be explained by a specific change of a driver such as surface wind stress change. The other is to divide upwelling (i.e., the vertical component of velocity) into upwelling along isopycnals or isopycnal upwelling (the vertical component of current velocity parallel to isopycnal surfaces) and upwelling across isopycnals or diapycnal upwelling (the vertical component of current velocity perpendicular to isopycnal surfaces) (Bryden and Brady 1985; Sloyan et al. 2003). It can be expected that the change in isopycnal upwelling is strong where thermocline flattening is large, but in order to determine the relative contribution of isopycnal and diapycnal upwelling, a quantitative analysis is needed.

The rest of this paper is organized as follows. Section 2 describes the climate models and method used in this study. Section 3a shows the relation between MME mean and intermodel variation for upwelling change under global warming. In section 3b, we divide upwelling into isopycnal and diapycnal upwelling components and examine the mechanisms of future changes of these components. Section 4 presents a summary and discussion.

2. Data and methods

The data used in this study are outputs of climate models participating in CMIP5 (Taylor et al. 2012), and were mainly obtained through the Program for Climate Model Diagnosis and Intercomparison (<http://cmip-pcmdi.llnl.gov/cmip5/>). We used the monthly data for the historical experiment up to December 2005 and those for the RCP8.5 emissions scenario from January 2006 to December 2100. In this scenario, greenhouse gas concentration and thus radiative forcing do not stabilize by 2100 (Liddicoat et al. 2013). We used outputs of 24 models, as listed in Table 1, and analyzed only the first ensemble of each model so as to treat all models equally.

Zonal and meridional components of velocity from CMIP5 outputs were estimated by a coordinate rotation

TABLE 1. CMIP5 models used in this study. (Expansions of acronyms are available online at <http://www.ametsoc.org/PubsAcronymList>.)

	Model	Institution	OGCM resolution (lon grid points $\times$ lat grid points $\times$ levels)
A	BCC-CSM1.1	Beijing Climate Center, China	360 $\times$ 232 $\times$ 40
B	CanESM2	Meteorological Administration, China Canadian Centre for Climate Modeling and Analysis, Canada	256 $\times$ 192 $\times$ 40
C	CCSM4	National Center for Atmospheric Research (NCAR), United States	320 $\times$ 384 $\times$ 60
D	CESM1-BGC		
E	CMCC-CESM	Centro Euro-Mediterraneo per i Cambiamenti Climatici, Italy	182 $\times$ 149 $\times$ 31
F	CMCC-CM		
G	CMCC-CMS		
H	CNRM-CM5	Centre National de Recherches Météorologiques, France	362 $\times$ 292 $\times$ 42
I	CSIRO-Mk3.6.0	CSIRO and Queensland Climate Change Centre of Excellence, Australia	192 $\times$ 189 $\times$ 31
J	FGOALS-s2	LASG, Institute of Atmospheric Physics, Chinese Academy of Sciences, China	360 $\times$ 196 $\times$ 30
K	GFDL CM3	NOAA/Geophysical Fluid Dynamics Laboratory, United States	360 $\times$ 200 $\times$ 50
L	GFDL-ESM2M		360 $\times$ 200 $\times$ 50
M	GISS-E2-R	NASA Goddard Institute for Space Studies, United States	288 $\times$ 180 $\times$ 32
N	HadGEM2-CC	Met Office Hadley Centre, United Kingdom	360 $\times$ 216 $\times$ 40
O	HadGEM2-ES		
P	IPSL-CM5A-LR	Institute Pierre-Simon Laplace, France	182 $\times$ 149 $\times$ 31
Q	IPSL-CM5A-MR		
R	IPSL-CM5B-LR		
S	MPI-ESM-LR	Max Planck Institute for Meteorology, Germany	256 $\times$ 220 $\times$ 40
T	MPI-ESM-MR		802 $\times$ 404 $\times$ 40
U	MRI-CGCM3	Meteorological Research Institute, Japan	360 $\times$ 368 $\times$ 51
V	MRI-ESM1		
W	NorESM1-M	Norwegian Climate Centre, Norway	320 $\times$ 384 $\times$ 70
X	NorESM1-ME		

using original velocity components in the directions of the model coordinate and coordinate angles relative to the meridian estimated from the coordinates. The upward component of velocity, which is not included in the standard outputs of CMIP5, was calculated from upward mass transport and the area of grid cell. All model outputs were interpolated to a common horizontal grid with $1^\circ \times 1^\circ$ resolution, and the three-dimensional data were further interpolated to a common set of 28 vertical levels with 13 levels in the top 400 m of the ocean.

To understand the mechanisms of change in MME mean, we conducted two analyses. The first was an analysis of the relation between MME mean and intermodel variation base on a linear regression. It is useful to illustrate here how this analysis works. Consider a case where a linear mechanism under consideration plays an important role in intermodel variation as well as the MME mean change. In this case, epoch differences, denoted by Δ , of explanatory variable X and response variable Y among models can be related by

$$\Delta Y_m = a\Delta X_m + \Delta Y_m^o + \Delta \varepsilon_m, \quad (1)$$

where subscript m represents the model number, a is a constant, ΔY_m^o indicates the collective contribution of other mechanisms than the first term, and ε represents variation unrelated either to ΔX_m or ΔY_m^o (i.e., “noise”). The MME change is thus expressed by

$$\Delta Y_{\text{MME}} = a\Delta X_{\text{MME}} + \Delta Y_{\text{MME}}^o. \quad (2)$$

This relation corresponds to the regression line with slope a and intercept of ΔY_{MME}^o . Therefore, the fractional contribution ratio of mechanism of the first term, denoted R , for the total MME change can be expressed as

$$R = \frac{a\Delta X_{\text{MME}}}{\Delta Y_{\text{MME}}}. \quad (3)$$

If the signs of the two terms of the right-hand side of Eq. (2) are the same (i.e., they work constructively), the ratio of Eq. (3) is always positive and between zero and unity (0%–100%). If the signs are opposite, there are two further cases. If the mechanism is under consideration is dominant, then the contribution ratio becomes larger than unity (>100%). If the considered mechanism is minor,

FIG. 1. Schematic scatter diagram between epoch differences of explanatory variable X and response variable Y among climate models. The dots indicate respective model changes and the plus symbol (+) denotes the change in MME mean; ΔX_{MME} , ΔY_{MME} , and a represent the changes in MME mean in X and Y , and the regression slope, respectively. The black line indicates the regression line. The blue arrow indicates the ΔY_{MME} explained by $a\Delta X_{\text{MME}}$. The contribution ratio is defined by the ratio between $a\Delta X_{\text{MME}}$ and ΔY_{MME} (see text).

then the contribution ratio becomes negative. In a scattered plot (Fig. 1), the contribution ratio, Eq. (3), is expressed by the distance between ΔY_{MME} and ΔY value at the intercept (blue arrow in the figure) relative to ΔY_{MME} itself.

It is apparent that this argument holds even if two or more linear mechanisms are at work, provided they are independent of each other. On the other hand, if one examines two mechanisms and finds their contributions are highly correlated, one cannot separately evaluate the contributions of respective mechanisms.

The second analysis is the division of total upwelling w_t into isopycnal upwelling w_i and diapycnal upwelling w_d , following Pedlosky (1996). Since an isopycnal upwelling is a vertical velocity component derived from isopycnal flow, one can obtain isopycnal upwelling from the following advection equation:

$$w_i \equiv u \frac{\partial h}{\partial x} + v \frac{\partial h}{\partial y}, \quad (4)$$

where h is the depth of isopycnals (positive upward), u and v are the zonal and meridional velocities, respectively, and w_i is the isopycnal upwelling velocity. The diapycnal upwelling is calculated as the difference

FIG. 2. Differences in upwelling along the equator (colors) between 1971–2000 and 2071–2100 for the 24-model MME mean under RCP8.5. Contours indicate the mean upwelling for the period 1971–2000. Stippling indicates regions where projected changes are statistically significant according to a Wilcoxon signed-rank test with a 5% significance level and more than 90% of the models agree on the sign of the change.

between the total upwelling and the isopycnal upwelling, that is,

$$w_d = w_t - w_i, \quad (5)$$

where w_t and w_d are the total and diapycnal upwelling velocities, respectively. The concepts of isopycnal and diapycnal upwelling have been introduced by previous studies of equatorial circulations and the Pacific cold tongue (Bryden and Brady 1985; Sloyan et al. 2003), but to the authors' knowledge they have not previously been employed in studies of future upwelling changes. Please note that the terminology of total upwelling is used for this analysis, and it is the same as the upwelling in the other analyses.

3. Results

a. Patterns of upwelling change

From 1971–2000 to 2071–2100, the MME mean upwelling decreases in the equatorial Pacific and Atlantic at a depth of about 0–200 m and increases in the eastern Indian Ocean at a depth of about 0–100 m (Fig. 2). The equatorial Pacific has the largest upwelling decrease among the tropical oceans. The Pacific upwelling decreases most strongly around 120°W at a depth of 100 m, by about 30% of its climatology, and this vertical velocity change is 3 times larger than that in the Atlantic. The MME mean upwelling strongly decreases in the eastern Pacific basin at a depth of 250 m or shallower, where all models agree on the sign of the change, and the

amplitude of the upwelling change is statistically significant according to a Wilcoxon signed-rank test at the 5% significance level (stippled regions in Fig. 2). At depths between 50 and 75 m, a relatively strong reduction in MME mean upwelling appears in the central Pacific, but only 70% of the models agree on the sign of the change. In the western Atlantic, 22 of the 24 models show a reduction in upwelling at depths of 25–150 m. At deeper depths than 150 m, the MME mean upwelling changes only a little. In the Indian Ocean, changes in upwelling are generally insignificant. These results are consistent with those of Vecchi and Soden (2007), who analyzed the outputs of CMIP3 models.

It is interesting to examine future upwelling changes in each model and how the changes are related to the reproductivity of upwellings in climate models. We compare the mean climatology of climate model in the twentieth century with the outputs of a $0.1^\circ \times 0.1^\circ$ high-resolution ocean model, the Ocean Model for Earth Simulator (OFES), forced by the wind stress estimated from QuikSCAT satellite from January 1999 to December 2008 (Sasaki and Nonaka 2006). For the Pacific Ocean, climate models generally reproduce the similar spatial structure of upwelling to that simulated by OFES, although weak upwellings deeper than 150–200 m in OFES between 180° – 160° W and around 120° W are not found in climate models (Fig. 3). Consistent with Fig. 2, most of models exhibit future weakening of upwelling in the eastern Pacific, but models do not agree well in the central Pacific, where upwelling is enhanced in some models.

Differences of mean climatological upwelling in climate models compared with OFES are larger for the Atlantic Ocean than for the Pacific Ocean (Fig. 4). About half the models substantially underestimate mean upwelling in the Atlantic. The models that have weak climatological upwelling also exhibit small future change, and thus these small epoch changes may be underestimated associated with the bias in the mean climatology. To get an idea about the influence of underestimating models, we recalculate MME upwelling change for the Atlantic without 11 models (out of 24) that have smaller climatological upwelling by 30% compared with OFES along the equator over 20° – 40° W and at depths of 20–150 m. The resultant MME upwelling change is enhanced by about 20%–30% from the upwelling shown in Fig. 2 in the Atlantic with essentially the same spatial pattern (not shown).

Some models from the same modeling center exhibit similar climatology and epoch differences both for the Pacific and Atlantic Oceans. For example, model pairs provided by the same modeling center—C and D, K and L, N and O, P and Q, and W and X (as labeled in Table 1)—show similar spatial structures.

b. Causes of upwelling changes

We first estimate how much of the intermodel variation in upwelling change can be explained by wind stress and stratification changes for the Pacific Ocean. Here, we also examined the influence of stratification change (Fig. 5b) because increased stratification has weakened upwelling in the current climate (e.g., Fiedler 2002). The intermodel epoch differences of equatorial upwelling are strongly correlated to epoch differences in zonally averaged zonal wind stress over the eastern Pacific from near the surface to 200-m depth (Fig. 5a). Therefore, we suggest that the intermodel variation in upwelling reduction around the EUC core and at shallower depths in the eastern equatorial Pacific is caused by intermodel differences in the weakening of the trade winds in that region. This is also true for the relatively shallow depths (50–100 m) in the central Pacific: the upwelling changes across models are well correlated by local (180° – 140° W) trade wind weakening (not shown), although for this region the models disagree on the sign of the upwelling change as mentioned above. The correlations between the changes in upwelling and changes in stratification are significant in some regions (Fig. 5b) but do not substantially overlap with the strong MME mean changes in upwelling.

The significant MME upwelling change in the equatorial eastern Pacific but insignificant in the central Pacific is related to the significance of MME change of the zonal wind stress (Fig. 6). The equatorial zonal wind stress changes statistically significant in the eastern Pacific, where the MME mean change has its peak, but not in the central Pacific. In addition, wind changes are insignificant across the equatorial Atlantic Ocean (not shown).

The mechanism of the intermodel variation of changes in upwelling can also partly explain changes in MME mean. In the eastern Pacific, the local wind stress change well explains ($r = -0.60$) the intermodel variation in upwelling change near the sea surface (Fig. 7a), consistent with the strong negative correlation near the surface shown in Fig. 5a. About two-thirds of the upwelling reduction near the surface is explained by the weakened trade wind over the eastern Pacific basin, represented by the contribution ratio $a\Delta\tau_{\text{MME}}^x/\Delta w_{\text{MME}}$ of 0.67. Around the EUC core in the eastern Pacific, the models that show a strong reduction in upwelling exhibit strong weakening of the local trade wind ($r = -0.63$) (Fig. 7b), again consistent with Fig. 5a. Similarly, the ratio $a\Delta\tau_{\text{MME}}^x/\Delta w_{\text{MME}}$ is 0.46, indicating that nearly half of the MME mean upwelling reduction around the EUC core is explained by the weakened local trade wind. Consequently, locally weakened trade wind due to global warming explains not only part of the intermodel variation but also part of the changes in MME mean for a weakening of near-surface

FIG. 3. (a)–(x) Equatorial upwelling differences (colors) between 1971–2000 and 2071–2100 for each of 24 models under RCP8.5 in the Pacific Ocean, and (bottom right) climatological upwelling of OFES simulation from 1999 to 2008. The panel letters denote the respective models in Table 1. The contour interval is $0.5 \times 10^{-5} \text{ m s}^{-1}$ and blue contour indicates $2.0 \times 10^{-5} \text{ m s}^{-1}$.

upwelling and a reduction of subsurface maximal upwelling at depths of 75–200 m in the eastern equatorial Pacific.

In the western Atlantic, both a weakened trade wind and enhanced stratification are significantly correlated with a reduction in upwelling at depths of 20–150 m (Fig. 8), with the negative correlations covering the region of strong upwelling reduction. The area-averaged upwelling change at depths between 20 and 150 m in the western Atlantic is highly correlated across models with the basin average of change in equatorial zonal wind ($r = -0.85$) and also with the basin average of change in stratification ($r = -0.70$) (Fig. 9). However, the changes in zonal wind stress and the enhanced stratifications are strongly correlated in the Atlantic ($r = 0.60$), in contrast

to the quite small correlation in the Pacific ($r = 0.11$). This dependency between the change in trade wind and the change in stratification makes it difficult to distinguish the relative contributions of the two mechanisms to the reduction in upwelling in the Atlantic as explained in section 2.

c. Changes in isopycnal and diapycnal upwelling

To obtain a better physical explanation for the change in upwelling, we divided upwelling, which is referred to as the total upwelling in this subsection, into isopycnal and diapycnal upwelling, as explained in section 2. As given by Eq. (5), the change in total upwelling w_t (Fig. 2) is equal to the sum of the isopycnal upwelling w_i

Upwelling Diff., Atlantic, RCP8.5 2071-2100 minus 1971-2000

FIG. 4. As in Fig. 3, but in the Atlantic Ocean. The blue contour indicates $1.0 \times 10^{-5} \text{ m s}^{-1}$.

(Figs. 10a,c) and diapycnal upwelling w_d (Figs. 10b,d). The statistically significant change in isopycnal upwelling occurs at depths between 75 and 200 m in the eastern Pacific (Fig. 10a), where isopycnals are strongly tilted, as will be shown later, whereas statistically significant diapycnal change occurs at depths between 0 and 75 m in the central to eastern Pacific (Fig. 10b). In the equatorial Atlantic, isopycnal and diapycnal upwelling show very little significant change, due to the small amplitude of changes in MME mean relative to large intermodel variation (Figs. 10c,d). We will therefore focus our attention on the upwelling change processes in the equatorial Pacific, especially the eastern equatorial Pacific, where upwelling decreases most. Next, we examine how near-surface diapycnal upwelling and isopycnal upwelling at

deeper levels are related to changes in subsurface density and circulation.

The near-surface upwelling in the eastern Pacific should be related to the Ekman pumping. The weakened zonal wind stress in the equatorial eastern Pacific (Fig. 6) implies that the Ekman pumping, which is closely related to Ekman divergence across the equator, should be weakened in future, because vertically integrated Ekman transport solely depends on the wind stress at the chosen latitudes. This is consistent with the aforementioned strong correlation between the change in total upwelling and the change in zonal wind stress (Figs. 5a and 7a). Consequently, it is strongly suggested that the weakening of the near-surface total upwelling in the eastern Pacific, due to the diapycnal upwelling reduction, is closely

FIG. 5. Intermodel correlation coefficient (colors) between epoch differences in upwelling at the equator and (a) epoch differences in eastern Pacific basin-averaged zonal wind stress (140°–80°W along the equator) and (b) epoch differences in the basin-averaged stratification (160°E–80°W along the equator between surface and 100 m). The epochs are 1971–2000 and 2071–2100 under RCP8.5. Contours indicate epoch differences in upwelling in both panels.

related to the reduction of the Ekman pumping caused by the weakened trade wind.

The reduction in isopycnal upwelling, which dominates the changes in total upwelling between 75 and 200 m in the eastern Pacific, should be linked to the reduction of the along-isopycnal velocity associated with the zonal velocity and/or zonal isopycnal flattening. The future zonal velocity significantly reduces beneath the EUC core in the current climate, accompanied by zonal velocity enhancement above the core (Fig. 11a), suggesting an upward migration of the EUC. Shallowing of the EUC due to global warming was reported by Luo et al. (2015) and Sen Gupta et al. (2012), but these studies did not investigate the relation between the EUC and upwelling. For the other mechanism, namely isopycnal flattening, the isopycnals corresponding to the EUC core indicate a flatter zonal slope in the future than at present (Fig. 11b). Therefore, both zonal velocity

FIG. 6. Zonal wind stress differences along the equator between 1971–2000 and 2071–2100 for 24-model MME under RCP8.5. Dots indicate locations where projected changes are statistically significant by a Wilcoxon signed-rank test with a 5% significance level and where at least 90% of the models agree in the sign of change.

reduction and zonal isopycnal flattening can contribute to reduced isopycnal upwelling in the eastern equatorial Pacific.

The contributions of zonal velocity change and isopycnal slope change to the reduction of MME mean isopycnal upwelling at depths between 75 and 200 m can be calculated from a linear expansion of the advection equation [Eq. (4)]. We separate each dependent variable into mean and anomaly components as follows:

$$\Delta w_i \approx \Delta u \left(\frac{\partial \bar{h}}{\partial x} \right) + \bar{u} \Delta \left(\frac{\partial h}{\partial x} \right) + \Delta v \left(\frac{\partial \bar{h}}{\partial y} \right) + \bar{v} \Delta \left(\frac{\partial h}{\partial y} \right), \quad (6)$$

where Δ indicates the difference between the two epochs (2071–2100 minus 1971–2000), and an overbar indicates the mean value for the period 1971–2000. The third and fourth terms on the right-hand side can be ignored because these two terms are much smaller than the first and the second terms. The first and second terms represent the change in isopycnal upwelling associated with zonal velocity and isopycnal tilt changes, respectively (Figs. 12a,b). The change in isopycnal upwelling at depths between 75 and 200 m in the eastern equatorial Pacific (Fig. 10a) is largely accounted for by the change in zonal isopycnal tilt. This indicates that the weakening of MME mean isopycnal upwelling in the eastern Pacific is mainly induced by a zonal flattening of the east–west slope of the EUC.

The changes in upwelling due to the flattening of the EUC slope also largely explain the intermodel variation of changes in isopycnal upwelling in the eastern Pacific (Fig. 12c). The corresponding regression slope of 0.71 is 1.5 times larger than the slope of 0.47 for changes in upwelling due to the zonal velocity change. Consequently,

FIG. 7. (a) Scatter diagram of epoch differences of area-averaged upwelling (140° – 100° W at depths of 20–75 m along the equator) and zonal wind stress average over the same longitudinal extent among climate models. The epochs are 1971–2000 and 2071–2100 under RCP8.5. The plus symbol (+) denotes the MME, and letters denote the respective models (Table 1). The black line indicates the regression line. The blue arrow indicates the change in MME mean upwelling associated with the MME mean basin-averaged zonal wind stress change. The dashed line indicates the change in MME mean upwelling at the point where the regression line crosses the line of zero zonal wind stress change. The correlation coefficient and slope are also shown. (b) As in (a), but for area-averaged upwelling (140° – 100° W at depths of 75–200 m along the equator) and eastern basin-averaged zonal wind stress (140° – 80° W).

flattening of the zonal pycnocline slope plays a dominant role in the intermodel variation of change in isopycnal upwelling as well as the change in MME mean at depths between 75 and 200 m in the eastern Pacific.

The zonal tilt of the equatorial pycnocline for El Niño events is well explained by a reduced gravity model (e.g., Jin 1997), which may quantitatively account for the EUC flattening under increased stratification. We employ a linearized reduced gravity model for the change in isopycnal

FIG. 8. As in Fig. 5, but for the Atlantic. The averaged area of change in zonal wind stress and change in stratification is 60° W– 10° E along the equator, and the stratification is given by the density difference between the surface and 100 m.

tilt, which can be expressed by the changes in stratification and zonal wind stress:

$$\Delta \frac{dh}{dx} = \frac{1}{\rho_0 H} \Delta \left(\frac{\tau^x}{g'} \right) \approx \frac{1}{\rho_0 H} \frac{1}{g'} \Delta \tau_x + \frac{1}{\rho_0 H} \frac{\tau_x}{g'^2} \Delta \frac{1}{g'}, \quad (7)$$

where $\langle \tau^x \rangle$ is the zonally averaged zonal wind stress along the equator, g' is the reduced gravity, ρ_0 is the reference density, h is the total local depth of the pycnocline, and H is the constant background layer thickness. The first term on the right-hand side indicates the influence of the change in zonal wind stress, while the second term indicates the changes in the density difference between the surface layer and deeper layer. The approximation of the right-hand side of Eq. (7) and the term in the middle are almost identical ($r = 0.99$). The reduced gravity change examined here represents the influence of stratification change onto the thermocline slope, while the stratification change influence examined earlier (Fig. 5b) is for the total upwellings. We employ this model in the eastern equatorial

FIG. 9. (a) As in Fig. 7a, but for area-averaged upwelling in the Atlantic (40° – 20° W at depths of 20–150 m along the equator) and Atlantic basin-averaged zonal wind stress (60° W– 10° E). (b) As in (a), but for area-averaged upwelling (40° – 20° W at depths of 25–150 m along the equator) and basin-averaged stratification (60° W– 10° E along the equator between the surface and a depth of 100 m).

Pacific (140° – 100° W), where change in isopycnal tilt plays the dominant role in the change in isopycnal upwelling (Fig. 12b). The reference density and the layer thickness are set to 1025 kg m^{-3} and 75 m, respectively. Reduced gravity is calculated using the zonal mean (140° – 100° W) density between the surface and the pycnocline and that between the pycnocline and a depth of 1000 m.

To know whether the reduced gravity model can be applied to EUC flattening, we examine the relationship between $\Delta h/dx$ and $\Delta(\tau_x)$ for the first term of Eq. (7) and that between $\Delta h/dx$ and $\Delta 1/g'$ for the second term. Here, the isopycnal depth h is defined as the depth of a constant density that is defined as the zonally averaged density between 140° and 100° W at a depth of 75 m. The correlation for the first term is as high as 0.84, but that for the second term is very low ($r = 0.17$). Thus, only the first term (i.e., the change

in wind stress) plays an important role in the change of the pycnocline slope. The regression slope for the first term is close to that expected from the reduced gravity model; that is, the regression slope is larger by only 25% than the coefficient of the first term, $1/(\rho_0 H g')$, with the regression line almost crossing the origin. These results indicate that the EUC flattening is explained by the change in zonal wind stress, which is consistent with the reduced gravity model.

4. Summary and discussion

We investigated the change in equatorial upwelling and its mechanisms in the Pacific and Atlantic Oceans until 2100 under RCP8.5 by analyzing MME mean and differences among 24 CMIP5 models. Zonal and meridional components of current velocities were estimated from velocities in the original coordinate system by a coordinate rotation. The estimation of zonal and meridional velocities allowed us to investigate three-dimensional circulation change and to diagnose isopycnal and diapycnal upwelling.

In the eastern equatorial Pacific, the local zonal wind stress, which will be weakened in the future (Fig. 6), influences the MME mean (Fig. 2) and intermodel variation (Figs. 5a and 7) in the changes in upwelling via two mechanisms. One is the decrease in diapycnal upwelling near the surface (Fig. 10b), which is associated with the weakened Ekman pumping. The other is the decrease in isopycnal upwelling at depths of 75–200 m around the EUC core (Fig. 10a), induced by EUC flattening (Fig. 11b). These mechanisms of upwelling changes are at work for both the MME mean and intermodel variation, as summarized in Fig. 13.

In the equatorial Atlantic, for which climate models suffer from upwelling underestimation, the intermodel variation in the reduction in total upwelling at a depth of 25–150 m, where there is a local maximum of reduction in MME mean upwelling, is significantly related to both the weakened trade wind and enhanced stratification (Fig. 8). Both these drivers also explain the change in MME mean upwelling at depths of 25–150 m (Fig. 9). However, it remains unclear how much of the total upwelling changes is contributed by each driver because the changes in wind and stratification are not independent in the equatorial Atlantic. Very little significant change in isopycnal and diapycnal upwelling is projected due to the small amplitude of changes in MME mean relative to the large intermodel variation (Figs. 10c,d). Thus, the mechanisms of upwelling change in the equatorial Atlantic are not well identified by the present analyses.

The reduction in near-surface upwelling, which is dominated by diapycnal upwelling, in the eastern equatorial Pacific can impact the SST there. The response of equatorial SST to global warming is important for tropical

FIG. 10. Differences in isopycnal upwelling (colors) between 1971–2000 and 2071–2100 for 24-model MME under RCP8.5 in the (a) equatorial Pacific and (c) equatorial Atlantic. Contours indicate mean isopycnal upwelling for the period 1971–2000. Stippling indicates regions where projected changes are statistically significant according to a Wilcoxon signed-rank test with a 5% significance level and more than 90% of the models agree on the sign of the change. (b),(d) As in (a) and (c), respectively, but for differences in diapycnal upwelling.

climate conditions such as ENSO (e.g., Knutson and Manabe 1995; Meehl and Washington 1996; Liu 1998; Meehl et al. 2000; Zhang et al. 2008; DiNezio et al. 2009; Liu et al. 2017). In particular, stronger warming in the eastern equatorial Pacific than in the surrounding regions may lead to more frequent extreme or strong eastern Pacific El Niño events (Cai et al. 2014, 2018). A possible mechanism for this future change in SST in the equatorial Pacific is reduced upwelling in response to weaker trade wind (Zhang et al. 2008; DiNezio et al. 2009). Our results indicate that upwelling reduction near the surface is closely related to the local wind (Figs. 5a and 7a) and is mainly due to diapycnal upwelling (Fig. 10b). This suggests that changes in the local Ekman pumping rather than the basin-scale circulation change including EUC are important in SST change. Of course, diapycnal and isopycnal upwellings might not be totally separated in reality; for example, waters transported along isopycnal surface can be finally upwelled via diapycnal upwelling. Further studies, for example heat budget analyses, would be useful to evaluate quantitatively how upwelling change impacts SST under the global warming.

In the present study, we employed an approach in which we treat the ocean as the system of interest and atmospheric changes especially those in equatorial zonal wind stress as external forcing, but as introduced in section 1, the tropics is the region where vigorous air–sea interaction occurs including Bjerknes feedback. Therefore, to fully understand the air–sea coupled system under the global warming, it is also necessary to understand how the atmosphere responds to the oceanic changes. For example, the statistically significant zonal wind stress changes (Fig. 6) in the eastern equatorial Pacific may be influenced by oceanic changes, including the upwelling and its influence on the SST.

The reduction in total upwelling in the equatorial Pacific can impact several important biogeochemical processes. The tropical Pacific is a region of high marine primary production (Sutton et al. 2017), rapid outgassing of CO₂ (Feely et al. 1999), and intense subsurface oxygen minimum zones (Wyrki 1962), all of which are influenced by upwelling. Models project a robust reduction in net community production in the eastern equatorial Pacific in the twenty-first century (Bopp et al.

FIG. 11. Depth–longitude cross sections along the equator of (a) zonal velocity difference and (b) σ difference between 1971–2000 and 2071–2100 for 24-model MME under RCP8.5 (colors). Contours indicate the mean (a) zonal velocity and (b) σ for the periods 1971–2000. Stippling indicates regions where projected changes are statistically significant according to a Wilcoxon signed-rank test with a 5% significance level and more than 90% of the models agree on the sign of the change. Green and red lines in (b) indicate the isopycnals of averaged σ along the EUC core (zonal velocity maxima) over the entire Pacific basin for the periods 1971–2000 and 2071–2100, respectively.

2013). This reduction is probably related to the reduced nutrient supply to the euphotic layer associated with the weakened upwelling in this region. Furthermore, changes in the structure of equatorial upwelling also have important implications for the tropical Pacific oxygen minimum zone (OMZ) and nitrogen loss in the OMZ (Deutsch et al. 2014). Changes in the EUC can also alter the O_2 supply to the OMZ (Stramma et al. 2010; Shigemitsu et al. 2017), partly counteracting the changes in biological O_2 demand (Ito and Deutsch 2013). The effects of these biogeochemistry

FIG. 12. (a) Changes in isopycnal upwelling due to differences in zonal velocity along the equator between 1971–2000 and 2071–2100 for 24-model MME under RCP8.5 (colors). (b) As in (a), but isopycnal upwelling changes due to differences in zonal isopycnal tilt. Contours indicate the mean zonal component of isopycnal upwelling for the period 1971–2000 in (a) and (b). (c) Scatter diagram of epoch differences of area-averaged isopycnal upwelling (140° – 100° W at a depth of 100 m along the equator) and contributions of zonal velocity (blue) and zonal isopycnal tilt (red) among climate models. The epochs are 1971–2000 and 2071–2100 under RCP8.5. The plus symbol (+) denotes the MME and letters denote the respective models (Table 1). Correlation coefficients and slopes are shown in the panel.

processes are likely to depend on the distinct contributions of isopycnal and diapycnal mechanisms to the reduction in total equatorial upwelling.

The equatorial upwelling is important process in physical climate, marine biogeochemistry, and ecosystems. Further studies are needed for understanding upwelling mechanisms as well as their impacts. In addition, it is important to address upwelling biases in climate models. For these future studies, the outputs from CMIP6 (Eyring et al. 2016), the next generation of CMIP, should be important.

FIG. 13. Schematic diagram illustrating the mechanisms of upwelling reduction in the eastern equatorial Pacific from 1971–2000 to 2071–2100. The weakening of trade winds (blue arrows) in the eastern Pacific induces weakening of the Ekman pumping associated with the diapycnal upwelling near the surface and also causes the EUC flattening and resultant reduction in the isopycnal upwelling at depths of 75–200 m. The variables shown in the zonal cross section are the same as in Fig. 2.

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