The Andes and the Southeast Pacific Cold Tongue Simulation

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ABSTRACT: Current climate models significantly overestimate precipitation and temperature over the southeast Pacific. Previous studies show that the Andes influence the southeast Pacific climate, but because they are a narrow mountain range, these effects are not well represented with the smoothing in low-resolution climate models. We hypothesize that the elevation of the Andes in low-resolution climate models is too low, allowing too much airflow across the mountains. Using the Community Earth System Model (CESM) version 1.2.2 with a slab-ocean setting, we blocked the low-level airflow across the mountains by adjusting the height of the Andes. Our results show that using higher Andes in the model enhances the anticyclonic motion and increases the subsidence of air in the southeast Pacific. The evaporative cooling from the strengthened surface wind and the radiative cooling from the formation of low-level clouds lower the sea surface temperature (SST) in the southeast Pacific. The enhanced descent over the southeast Pacific inhibits precipitation. We also performed a sensitivity test, varying the elevation of the Andes from 0 to 6 km, and analyzed the resulting energy budget and climate in both hemispheres. An increase of the elevation of the Andes causes the energy flux equator (EKE) to shift northward, as a result of the decrease of the energy absorption in the Southern Hemisphere. Our research suggests that the Andes act as a barrier in maintaining the climate over the southeast Pacific Ocean, and this effect is not well represented in low-resolution climate models.

KEYWORDS: Orographic effects; Topographic effects; Atmosphere-land interaction; General circulation models; Model errors

1. Introduction

Changes in the tropical eastern Pacific climate have significant influence on global and regional climate systems (Xie et al. 2007). In the tropical Pacific, the easterly trade wind moves warm surface ocean water toward the western Pacific and forms the western Pacific warm pool. Upwelling of deep ocean water in the eastern Pacific lowers the sea surface temperature (SST), forming the eastern Pacific cold tongue. The ascending branch of the Walker circulation is located in the western Pacific warm pool region and is accompanied by heavy rainfall, while the descending branch is located in the eastern Pacific cold tongue accompanied by strong lower-troposphere stability and low-level clouds (Xie 1998). Ocean-atmosphere interaction in this region leads to the El Niño–Southern Oscillation (ENSO) phenomenon (Dommenget 2010), which influences temperature and precipitation of other regions (Leathers et al. 1991) by the Pacific–North America (PNA) teleconnection

pattern (Yeh et al. 2018). Furthermore, climate model studies suggest that the cooling in the eastern Pacific in the recent decade is responsible for the slowing down of the global warming trend (Kosaka and Xie 2013). Therefore, the correct simulation of the climate over this region is crucial.

Despite its importance, there is a large bias in simulating the eastern Pacific climate. Climate models have difficulty in simulating the eastern Pacific climate where low-level clouds are formed by large-scale descending air (Bony 2005). Analysis of climate models from phase 5 of the Coupled Model Intercomparison Project (CMIP5; Taylor et al. 2012) reveals that most models simulate too high SSTs and too much precipitation in the tropical southeastern Pacific (Wang et al. 2014). The overestimation of precipitation in the tropical South Pacific leads to the double intertropical convergence zone (ITCZ) problem (Takahashi and Battisti 2007b).

Previous studies have suggested that the SST and precipitation biases are closely related to the atmospheric energy budget in climate models (Hwang and Frierson 2013; Li and Xie 2014; Adam et al. 2016). Adam et al. (2018) and Boos and Korty (2016) argue that the asymmetric SST bias is related to the hemispherically and zonally uneven heating anomalies, possibly due to the errors in simulating the radiative effect of clouds and the strength of the Atlantic meridional overturning circulation (AMOC) (Wang et al. 2014). Overall, studies suggest that models simulate too little reflection of shortwave radiation (Williams et al. 2013), and as a result the southeast Pacific is too warm.

These above mentioned biases result in a more symmetric distribution of SST and precipitation over the Pacific Ocean in the Northern Hemisphere (NH) and the Southern Hemisphere (SH). However, average SSTs are higher in the NH than in the

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SH, and the ITCZ is located in the NH (Marshall et al. 2014). For SST, the asymmetry is probably related to the uneven distribution of continents and topography. Philander et al. (1996) suggests that the shape of South America is more favorable for upwelling, forming low SSTs in the southeast Pacific. Moreover, the relatively low SSTs in the South Pacific form a regional high surface pressure while the higher SSTs in the North Pacific form a low surface pressure. This asymmetric pattern brings an anticyclonic motion in the South Pacific and a cyclonic motion in the North Pacific. These motions strengthen the easterly trade winds in SH and weaken that in NH. The stronger trade winds in the SH further decrease SST with wind–evaporation–sea surface temperature (WES) feedback (Xie and Philander 1994).

For precipitation asymmetry, Takahashi and Battisti (2007a) argued that the Andes are the most important driver for the northward distribution of the eastern Pacific ITCZ. The Andes block the midlatitude westerlies from the Pacific Ocean, and squeeze the isentropic surface as low-level air approaches the Andes. To conserve potential vorticity, the midlatitude winds have to turn equatorward as they descend. This effect brings cold and dry air from the polar region and enhances the anticyclonic motion (Rodwell and Hoskins 2001) as well as the subsiding motion in the SH. This dynamic motion also affects the formation of clouds. Because the ITCZ is located in the NH, the southern branch of the Hadley circulation is stronger, enhancing the descent over the southeast Pacific region. The enhanced descent helps maintain boundary layer clouds (Mechozo et al. 1995) and temperature inversion above the cold tongue (Wang et al. 2004).

Several experiments have highlighted the importance of the Andes in the formation of the climate features (e.g., clouds and SSTs) and energy transport over the Pacific. Xu et al. (2004) removed the Andes in a regional climate model and showed that the removal of the Andes caused a significant reduction of stratocumulus clouds and decrease in absorption of solar radiation on the sea surface in the southeast Pacific. Feng and Poulsen (2014) changed the elevation of the Andes from 1 to 3 km and found that the uplift of the Andes accounts for half of the cool SST over the cold tongue and significantly strengthens the Walker circulation and cloud radiative cooling. In addition, using aquaplanet modeling experiments, Maroon et al. (2015) showed that adding the Andes did lower the SST over the southeast Pacific and cause a northward propagation of the ITCZ.

The Andes are very important for the maintenance of the SST and precipitation features in the southeast Pacific. However, because it is a narrow mountain range, its elevation may be too low to influence the southeast Pacific climate when the topography is smoothed in low-resolution models. As a result, it cannot effectively block the midlatitude westerlies. Xu et al. (2004) and Gent et al. (2010) showed that a higher-resolution climate model has a better performance in simulating the wind and temperature pattern, and this is mainly because of the better-resolved orography. Therefore, we hypothesize that the CMIP5 model bias in the southeast Pacific is related to the topography settings in climate models. We modified the topography setting in South America and examined the influence of the topographic smoothing of the Andes on the southeast Pacific

climate. Our results show that when the low-level wind is properly blocked by the Andes, SSTs in the southeast Pacific decrease as a result of enhanced wind and low-level clouds. The rest of the paper is organized as follows. Section 2 introduces the climate model and the model settings of our experiments. Section 3 compares the precipitation and SST of model results with control and higher Andes cases, as well as the climate sensitivity for the elevation of the Andes and the change in atmospheric energy transport. Section 4 summarizes our findings.

2. Methods

a. Model description

We used the National Center for Atmospheric Research (NCAR) Community Earth System Model (CESM) version 1.2.2. The Community Atmospheric Model version 4 (CAM4) is the atmosphere component of CESM. We ran the model with the finite-volume (FV) dynamical core and a slab ocean. In a slab ocean model, only the interaction between atmosphere and mixed layer of the ocean is considered, and the SST is determined by the depth of mixed layer and the energy fluxes (Lee 2019). Although a slab-ocean model ignores the change in the ocean circulation, it is widely used to study the response of the climate system to an imposed forcing because the atmosphere adjusts very fast to the environment. The ocean circulations take hundreds to thousands of years to reach an equilibrium (Neale et al. 2013), so a slab-ocean model can capture the fast response resulting from the atmospheric circulation.

To assess the skill of CESM-CAM4 and the effects of changing topographic setting of the Andes in simulating the southeastern Pacific, we compared our model results with instrumental records. For SST, we used the average of mean annual SST from 2004 to 2018 derived from the Hadley Centre Sea Ice and Sea Surface Temperature (HadISST) dataset ($1^\circ \times 1^\circ$) (Rayner et al. 2003). For precipitation, we used the average of mean annual precipitation over 1979–2018 from the Tropical Rainfall Measuring Mission (TRMM) for comparison with models (Kummerow et al. 1998).

b. Experiments

Compared with the 1-km high-resolution topography from the National Oceanic and Atmospheric Administration (NOAA) National Geophysical Data Center (NGDC) Global Land One-km Base Elevation (GLOBE) topography (Hastings and Dunbar 1999) (Fig. 1a), the $1.9^\circ \times 2.5^\circ$ resolution of the CESM control case has much smoothed topography (Fig. 1b). In our experiment, we take the highest elevation within each model grid point along the Andes, as shown in Fig. 1c, so that the Andes can effectively block the zonal winds. The model results for the original topography setting (Fig. 1b) will be referred to as the control run, and the output for the modified elevation (Fig. 1c) will be referred to as the “Higher Andes” case. To compare with modern satellite data, the CO_2 concentration is set to be 400 ppm.

To examine how different elevations of the Andes affect the southeast Pacific climate, we also ran a series of sensitivity tests for the elevation of the Andes. We set the elevation of the

FIG. 1. Elevation in South America from the (a) National Oceanic and Atmospheric Administration (NOAA) National Geophysical Data Center (NGDC) Global Land One-km Base Elevation (GLOBE) topography, (b) $1.9^\circ \times 2.5^\circ$ resolution of CESM control run, and (c) Higher Andes topography. The area where we performed the sensitivity test is marked with a red line.

region within the red line in Fig. 1c to 50 m and to 1, 2, 3, 4, 5, and 6 km.

3. Results and discussion

a. Change in SST

Compared with the observations, simulated SST in the control run is too high in the Pacific, particularly in the northern and southern tropical/subtropical Pacific between 20°S and 20°N

(Fig. 2d). In the North Pacific, the overestimation is most severe near 20°N and 120°W on the west coast of North America with the magnitude of 3°C , and it expands toward the central Pacific. In the South Pacific, the SST bias is highest near the west coast of South America and also expands toward the central Pacific. This bias is about 2°C on average. The regions where SST is overestimated are in the region from 160°E to 80°W in the South Pacific over the tropics, which means that the extent of the cold tongue in the South Pacific is underestimated. In the

FIG. 2. (left) Annual mean sea surface temperature from (a) HadISST, (b) control CESM, and (c) Higher Andes CESM. (right) SST difference between (d) control and HadISST, (e) Higher Andes and HadISST, and (f) Higher Andes minus control. Boxes 1 to 4 in (f) are the areas where we did the energy budget analysis.

TABLE 1. RMSE for SST. The tropical Pacific is the ocean region covering 40°S–40°N and from 120°E–60°W.

SST RMSE (°C)	Tropical Pacific	NH tropical Pacific	SH tropical Pacific
Control	1.52	1.37	1.66
Higher Andes	1.29	1.28	1.29

midlatitude area of the northwest Pacific, the SST is abnormally cold compared with the observation (Fig. 2d). The change in the Atlantic Ocean and Indian Ocean is not as obvious as in the Pacific Ocean (see Fig. S1 in the online supplemental material), so it is not shown in the main text.

The warm SST bias in the South Pacific in the control run is reduced in the Higher Andes case. (Figs. 2e,f). In 0°–20°S, the SST difference decreases from 1.47°C to less than 0.40°C. It also induces the Pacific–North American teleconnection pattern (PNA) in the North Pacific (Fig. 2f), with positive SST anomaly over the mid- and high-latitude North Pacific and negative SST anomaly over the tropical Pacific, similar to the results from Kosaka and Xie (2013). This teleconnection explains how the cooling over the southeast Pacific SST also eliminates the SST bias over the tropical North Pacific to some extent, by triggering Rossby waves that transport a cold SST signal in the tropical area to the subtropical regions as a warm SST signal (Dai et al. 2017).

To quantitatively show the change due to the modification of the Andes, we calculated the root-mean-square error (RMSE) between the model outputs in each case and the HadISST (Table 1). In the tropical North Pacific, the RMSE has been reduced by 6.5%, while in the tropical South Pacific it has been reduced by 22.3%. The overall RMSE over the tropical Pacific is reduced by 15.1%.

Our results suggest that the evaporative cooling and radiative cooling reduced the bias in the southeast Pacific SST (see below). Evaporative cooling accounts for the lowering of SST in the central South Pacific, while the radiative cooling explains the change in coastal areas of South America.

The evaporative cooling comes from stronger winds. In the tropics, the zonal wind blows from east to west. Near the west coast of South America, the strong northward wind induces the Ekman transport away from the coast of surface ocean water, resulting in low SSTs in the southeast Pacific. Around 120°W in the South Pacific, the low-latitude easterlies, midlatitude westerlies,

and equatorward wind on the coast of South America form an anticyclonic motion, establishing a high surface pressure center in the southeast Pacific (Fig. 3a).

By increasing the elevation of the Andes, the northward motion near the west coast of South America, as well as the easterlies near 20°S, is enhanced (Fig. 3c). The stronger northwestward surface airflow induces evaporation over the ocean, increases latent heat flux, and results in a lower SST in the central South Pacific. We compared the surface wind distribution with the ECMWF reanalysis data (Fig. S2), but it is difficult to tell if there is an obvious improvement due to the large discrepancy.

The radiative cooling comes from the increase of low-level clouds. Xu et al. (2004) found that the narrow steep Andes has a significant contribution to the formation of clouds over the southeast Pacific, and therefore results in the latitudinal asymmetry in the eastern Pacific climate. In our experiment, the change in the coastal southeast Pacific climate is the result of the change in low-level cloud fraction. The Higher Andes setting in the model enhances the descent motion of air over the subtropical ocean, forming a low-level trade inversion and low-level marine boundary layer (MBL) clouds (Fig. 4a) (Bony 2005). This region covers the longitude from 120°W to 80°W in subtropical areas of the southeast Pacific. The radiative effects of low clouds decrease local SSTs.

Figures 4b and 4c show the change in medium and high cloud fraction. Both medium and high cloud fractions show a decrease in the same region where SST is lowered (Fig. 2f). The decrease of medium and high clouds ends up with a weaker warming radiative effect. Medium and high clouds show an increase in the southwest Pacific where precipitation bias in the South Pacific convergence zone (SPCZ) is improved (section 3b).

To examine the radiative effect from the change of cloud fraction, we show the distribution of energy budget at the surface (Figs. 4d–f). Therefore, we examine the distribution of the change in shortwave radiation (Fig. 4d), outgoing longwave radiation (Fig. 4e), and change of energy absorption at the surface (Fig. 4f). Our results show that with the higher Andes, the absorbed solar radiation shows a decrease near the west coast of South America at the latitude near 20°S where low-level clouds increase. From the equatorial western Pacific to the southeast Pacific, both incoming shortwave energy and outgoing longwave energy increases as a result of the decreasing medium and high cloud fractions in this area. However, the incoming shortwave

FIG. 3. Surface latent heat (color contours) and horizontal wind at 850 hPa (arrows) for (a) the Control run and (b) the Higher Andes run, and (c) their difference. Box areas are as in Fig. 2.

FIG. 4. Changes in cloud fraction from the Higher Andes to Control cases, for (a) low-level (surface pressure to 700 hPa), (b) medium-level (700 to 400 hPa), and (c) high-level clouds (400 hPa to the model top). Also shown are changes in surface radiation: (d) incoming shortwave, (e) outgoing longwave, and (f) net radiation at the surface. Box areas are as in Fig. 2.

radiation increases by a great amount and the increase of outgoing longwave radiation only offsets a small part of it. As a result, the overall change of the surface radiation still shows an effect of heating up the major areas of the South Pacific, and cooling down the coastal area of South America.

To analyze how much different types of energy accounts for the decrease in the SST under the modification of the Andes topography, we took four boxes as shown in Fig. 2 where SST is decreasing and calculated the average energy change within each box (Table 2). In the equatorial western Pacific (box 1) and central South Pacific (box 3), the surface is absorbing more shortwave radiation, but it is balanced by the increased outgoing longwave radiation and stronger evaporation. In the coastal region of South America (box 4), the evaporation is getting smaller, and the incoming shortwave radiation has decreased by a greater amount. For the region over the tropical western Pacific (box 2), the high-level cloud fraction has reduced with the Higher Andes setting (Fig. 4c), increasing upward longwave radiation and decreasing SST.

However, the change in cloud fraction is more likely to be a feedback of the change in the SST instead of directly causing SST change (Takahashi and Battisti 2007a). Thus, it is necessary to diagnose what would be the most initial change that is made by the modification of the Andes topography. Following Takahashi and Battisti (2007a), we used the same equation as

$$\Delta T = \alpha_r \left[-\frac{\Delta U}{U_r} + \frac{1}{1 + B_r} \frac{\Delta q}{q_{sr} - q_r} + \frac{B_r}{1 + B_r} \frac{\Delta T_a}{T_r - T_{ar}} - \frac{\Delta F}{(1 + B_r) LHF_r} \right], \quad (1)$$

in which

$$\alpha_r = (1 + B_r) \left[\frac{\gamma}{LHF_r} + \frac{1}{q_{sr} - q_r} \left(\frac{\partial q_s}{\partial T} \right)_r + \frac{B_r}{T_r - T_{ar}} \right]^{-1} \quad (2)$$

and B is the Bowen ratio, which equals SHF/LHF , and the subscript r indicates the control case. Also, U is the surface wind speed, q is the specific humidity near the surface, and q_s is the near-surface saturation specific humidity; T is SST and T_a is near-surface air temperature; ΔF is the extra radiative forcing.

The four terms in Table 3 represent the change in SST due to the change in surface wind speed (U term), specific humidity (q term), surface air temperature (T term), and radiative fluxes (F term). Here, change in shortwave radiation is only considered as a feedback, and ΔF is estimated as a change in longwave radiative forcing on the SST due to the change in upper-tropospheric humidity with the elevation of the Andes (Takahashi and Battisti 2007a). To know the relative importance from the change in these parameters, we are only comparing these four terms inside the parentheses. Table 3 shows the values of these four terms in the box shown in Fig. 2.

It is clear that in the central and eastern Pacific region (boxes 2–4), the most important contribution to the decrease of SST comes from the decrease of specific humidity (q) term. This implies that in the equatorial eastern Pacific, central South Pacific, and southeast Pacific area, the lowered SST in the Higher Andes run results from the drying effect associated with the lifting of the Andes. The strengthening of the anticyclonic motion brings the dry air from the upper atmosphere and induces stronger evaporation over the entire South Pacific. However, when SST decreases, the saturation vapor pressure will decrease according to the Clausius–Clayperon equation.

TABLE 2. Area-averaged deviations in SST ($^{\circ}C$) and net surface SWs, LWs, LHF, and SHFs ($W m^{-2}$) for the Higher Andes minus Control run. Fluxes are positive into the ocean. See text for the location of the boxes.

Box	ΔSST	ΔSW	ΔLW	ΔLHF	ΔSHF
1	−0.63	15.16	−4.13	−10.64	−0.24
2	−0.77	−0.02	−2.1	3.97	−0.58
3	−1.01	21.43	−8.02	−11.73	−0.51
4	−1.77	−16.23	4.62	13.61	−0.59

TABLE 3. Area-averaged relative contributions to SST changes, through latent and sensible heat fluxes, by changes in surface wind speed (U term), specific humidity (q term), surface air temperature (T_a term), and longwave radiative fluxes (F term).

Box	U term	q term	T_a term	F term
1	-0.16	-0.17	-0.02	-0.03
2	0.04	-0.26	-0.02	-0.02
3	-0.10	-0.33	-0.04	-0.05
4	-0.04	-0.31	-0.09	0.03

This reduction in specific humidity is the strongest negative feedback that prevents evaporation. Even though we are not considering the change in the shortwave radiation as the force in the equation that initiates the lowering of the SST, it is the major feedback that enhances the cooling. The colder sea surface temperature induces the formation of low-level clouds by turbulent mixing in the atmospheric boundary layer (Schneider et al. 2019). The low-level clouds blocking the insolation is the most important factor that enhances this cooling effect and further lowers the SST (Takahashi and Battisti 2007a). In the tropical western Pacific region represented by box 1, the increase of near-surface wind speed and the drying effect play similarly important roles in the decrease of the SST. In all the four boxes, the longwave radiative effect (F) term is relatively smaller than the other terms, implying that the change in the longwave radiation is not an important original cause of the lowered SST.

b. Change in precipitation

In the control run (Fig. 5b), the SPCZ is much stronger and parallel to the equator compared with observations (Fig. 5a),

and it covers the area from 120°E to 120°W. The control run fails to predict the strength and the location of precipitation observed over the South Pacific (Fig. 5d). The SPCZ shifts northeastward and is much stronger in the control run than in the observation. As a result, the cold tongue only extends to 120°W in the South Pacific, and the dry band at the equatorial eastern Pacific is too narrow.

When we increase the height of the Andes, the narrow, dry band at the equatorial eastern Pacific strengthens and extends westward, and the triangular dry region in the southeast Pacific extends to 140°W (Fig. 5c). The location of the SPCZ is closer to the observation in the Higher Andes run (Fig. 5e), and the strong precipitation bias in the central South Pacific is somewhat eliminated. The model results indicate that the higher-elevation setting in the Andes can provide a better simulation of the dry zone in the southeast Pacific. The change in precipitation over other ocean basins are not as significant (Fig. S3).

The RMSE is also calculated for precipitation with reference to the TRMM data (Table 4). The RMSE for the tropical South Pacific has significantly reduced (by 29.2%). Even though the RMSE is slightly increased in the tropical North Pacific, the RMSE for the entire tropical Pacific is still reduced by 9.29%. It proves that the simulation of the tropical Pacific precipitation is getting better with the Higher Andes setting.

The vertical motion at 500 hPa is shown in the line contours in Fig. 5. The descending motion is coupled with divergence and the ascending motion is coupled with convergence. Because of the westward propagation of the convergence center in the Higher Andes run, the descent motion in the midlatitude southeast Pacific is enhanced (Fig. 5f) along with the enhancement of equatorward wind (Fig. 3c). This is because when airflow is blocked by the

FIG. 5. (left) Annual mean precipitation from (a) TRMM, (b) control CESM, and (c) Higher Andes CESM (colors). (right) Precipitation difference between (d) control and MODIS, (e) Higher Andes and MODIS, and (f) Higher Andes and control. Contours show vertical velocity at 500 hPa (Pa s^{-1}), with solid lines as positive values (descending motion) and dashed lines as negative values (ascending motion).

TABLE 4. RMSE for precipitation. The tropical Pacific is the ocean region from 40°S to 40°N and from 120°E to 60°W.

Precipitation RMSE (mm day ⁻¹)	Tropical Pacific	NH tropical Pacific	SH tropical Pacific
Control	1.83	1.70	1.95
Higher Andes	1.66	1.89	1.38

higher Andes, the midlatitude westerly wind will be redirected toward the equator. This equatorward motion not only brings cold and dry air from higher latitudes, but also descends as a result of the conservation of potential vorticity. This descent motion inhibits convection in the southeast Pacific, resulting in a reduction in precipitation over the South Pacific Ocean.

c. Regional averaged pattern

Figure 6 shows the regional distribution of SST and precipitation. In the zonal distribution of SST over the South Pacific (Fig. 6a), the modeled SSTs are higher than the observed SSTs across the whole Pacific basin, with a smaller bias in the western Pacific and a larger bias in the eastern Pacific. In the eastern Pacific, the modeled SST is 3°C higher than the observation, demonstrating that the simulated southeast Pacific cold tongue is too weak. With the higher elevation of the Andes, the eastern Pacific SSTs become similar to the observation. In the western Pacific, however, the SST bias in the Higher Andes case is 0.8°C, somewhat higher than the control run (0.6°C).

In the meridional distribution of the SST (Fig. 6b), the SSTs in control run are about 1°C higher than observation throughout the whole tropical area. With the higher Andes, this difference has been reduced by half, and the results are more similar to the observation, particularly near the equator. This improvement is related to the reduction of incoming solar radiation at the surface due to the increase in low clouds and the increased surface evaporative cooling induced by the stronger winds.

Figure 6c shows the zonal distribution of precipitation in the SH tropical region in the Pacific area. In observation, precipitation decreases almost linearly from west to east in 160°E to 90°W. However, in the control run, the precipitation values are high from 180° to 130°W with a peak at 130°W, where simulated

precipitation is twice as high as the observation. With the higher Andes, precipitation over the central Pacific Ocean decreases from 5 to 3.5 mm day⁻¹. Although it is still higher than the observation, the precipitation peak shifts westward to 150°W and is weakened. The higher elevation in South America deflects the midlatitude wind to turn counterclockwise, inhibiting precipitation off its west coast, and enlarges the dry and cold area, resulting in the westward migration of precipitation center.

Figure 6d is the meridional distribution of precipitation over the Pacific region. The observations show a large contrast of precipitation between the North and South Pacific. However, both the Control and the Higher Andes runs clearly show obvious double precipitation centers on both hemispheres. Results from the Control run are similar to the observed values from 2°S to 8°N, but are highly overestimated in the SH. Results from the Higher Andes case exhibit a slight decrease of precipitation over the region from 20°S to 9°N, but the precipitation values near the equator and in the North Pacific are underestimated.

To understand how the elevation of the Andes changes the east–west circulation over the Pacific, we calculate the streamfunction of divergent component of zonal wind over the equatorial Pacific (Fig. 7). The result shows that the Walker circulation is strengthened with the higher elevation of the Andes. This result agrees well with the improved east–west temperature gradient and the larger cold tongue area.

d. Sensitivity test and energy budget analysis

To obtain more insight into how the elevation of the Andes affects the climate over the southeast Pacific Ocean, we performed a series of sensitivity tests. In the area outlined by red lines in South America (Fig. 1c), we set the elevation to 50 m (indicated as the “No Andes” run in the following sections) and 1, 2, 3, 4, 5, and 6 km.

Figure 8 presents the zonal and meridional distributions of SST and precipitation of the cases with different heights of the Andes. In the SST zonal distribution (Fig. 8a), the SSTs do not show an apparent difference with different heights of the Andes on the western Pacific at the longitude from 160°E to 170°W. The differences in SSTs increase eastward. By changing the elevation in the Andes area from 50 m to 6 km, the

FIG. 6. (a),(c) Zonal and (b),(d) meridional distribution of annual mean (a),(b) SST and (c),(d) precipitation. SST is calculated only over the ocean area, and precipitation includes both land and ocean.

FIG. 7. Streamfunction of divergent component of zonal wind averaged over 5°S to 5°N for the Pacific region. Streamfunction is calculated as $\Psi = a \Delta \phi \int_0^p u_D dp/g$ as in Yu et al. (2012). Because the Walker circulation and the Hadley circulation are interconnected, the change in the vertical motion can result from both the zonal and meridional circulation. Therefore, we use the Ψ -vector method of Keyser et al. (1989) to partition the irrotational part of the three-dimensional flow into a pair of orthogonal two-dimensional circulations.

differences in SSTs over the southeast Pacific can be as large as 7°C. In the SST meridional distribution, with the increasing height of the Andes (Fig. 8b), the overall tropical SSTs over the Pacific basin decrease, and the greatest change occurs at around 7°S. The decrease of SST over the NH is much weaker than over the SH, resulting in a meridional asymmetry SST pattern over the Pacific Ocean.

Similar to SST change, in the zonal precipitation distribution (Fig. 8c), as the height of the Andes increases, the precipitation in the western Pacific does not change. However, from 160°E to 90°W, it decreases almost linearly to the east. With lower elevations of the Andes, precipitation peaks at 130°W over the South Pacific. With the increasing elevation of the Andes, this precipitation peak is lowered and migrates westward. When the elevation is as high as 5 or 6 km, the precipitation pattern is very similar to the observation, although this elevation is unrealistic. However, we can still conclude that the Andes with higher elevation does a better job in inhibiting precipitation over the east and central South Pacific. In the meridional precipitation distribution (Fig. 8d), when the Andes is removed, the precipitation pattern is very symmetric about the

equator and shows the double ITCZs. The increasing elevation of the Andes significantly decreases the precipitation in the SH, while it only decreases the precipitation in the North Pacific by a small amount. As a result, the north–south precipitation difference increases. With the increasing elevation of the Andes, there is less precipitation along the equator, consistent with the low SST and increasing amount of low-level clouds.

The Control run (solid black lines) model results show a similar distribution as the case with a 2-km elevation of the Andes, while the Higher Andes run (dashed black lines) results show a distribution similar to the case of 4 km. These results suggest that 2 and 4 km are the effective heights of the Andes in the Control and the Higher Andes cases. This explains why in the Control run, the height of the Andes is not enough to block the zonal winds.

To have a more straightforward comparison for the sensitivity test results, we plotted Fig. 9 as another interpretation for Figs. 6 and 8. From the sensitivity test (gray line), the SST over all the areas in the Pacific, and the precipitation over the South Pacific (Figs. 9a–g shows a decreasing trend with the increasing height of the Andes. The Control run (red line) overestimates

FIG. 8. (a),(b) Sensitivity test for the height of the Andes and SST and (c),(d) sensitivity test for the height of the Andes and precipitation.

FIG. 9. Sensitivity test comparison and results for (a)–(d) SST and (e)–(h) precipitation: (a),(e) Change in the southeast (SE) Pacific, 0° – 20° S, 100° – 120° W; (b),(f) change in the southwest (SW) Pacific, 0° – 20° S, 160° E– 180° ; (c),(g) change in the south-central (SC) Pacific, 5° – 15° S, 180° – 120° W; and (d),(h) change in the north-central (NC) Pacific, 5° – 15° N, 180° – 120° W.

the SST for all regions over the Pacific compared with the observation (black line), and also overestimates the precipitation except for the southwest Pacific. In the Higher Andes case (blue line), the simulation result approaches the observation obviously in the SST and precipitation over the southeast Pacific and the south-central Pacific. Although there is slight increase of biases in the southwest Pacific SST and the north-central Pacific precipitation, and the southwest Pacific precipitation barely changes with the Higher Andes setting, this modification can be still considered effective.

It is also worth noting that the observation values for precipitation do not fall on the sensitivity test lines at all in the precipitation figures (Figs. 9e–h). This indicates that changing the height of the Andes cannot entirely solve the precipitation biases in the CESM. This precipitation bias, which is related to the ITCZ problem, is a result of atmospheric net energy input biases near the equator and the cross-equatorial divergent atmospheric energy transport (Adam et al. 2018; Xiang et al. 2017).

To achieve further insight into this problem, we analyze meridional energy transport in the atmosphere (Fig. 10), a factor that is shown to be closely related to the location of the ITCZs (Bischoff and Schneider 2016). Adam et al. (2018) suggested that ITCZs are located farther south compared with the observation, and this bias in the ITCZ is related to the

atmospheric energy budget. They suggest that the southward shift of the ITCZ is related to either too much energy absorption in the Southern Hemisphere or too little energy absorption in the Northern Hemisphere. Here, we used the energy flux equator (EFE), a latitude where northward energy flux becomes zero, to indicate the location of the ITCZ. EKE shifts northward as the elevation of the Andes increases, and all the cases show positive slope and a single intersection at the zero line (Fig. 10a). Figure 10b shows that more energy is transported by the atmosphere from NH to SH with the higher Andes, and the observation value is consistent with the model result when the height of the Andes is between 5 and 6 km. These two figures indicate that with the increasing elevation of the Andes, the global mean ITCZ migrates northward. Since the Andes is an important factor that shapes the north–south ITCZ asymmetry (Takahashi and Battisti 2007a), correcting the effective height of the Andes is necessary to improve the global energy budget and large-scale circulation.

4. Summary and conclusions

Eastern Pacific climate has an important regional and global influence (Collins et al. 2010). However, climate models have difficulty in simulating the cold tongue off the west coast of South America. For example, compared with observations,

FIG. 10. (a) Northward atmospheric energy transport (AET) near the equator. (b) Cross-equator southward energy transport.

CESM overestimates temperature and precipitation in the southeast Pacific region (Bogenschutz et al. 2018).

In this study, using the CESM slab ocean model, we showed that the smoothed topography in South America will influence the model's simulation for precipitation and SST over the southeast Pacific. The Andes is a narrow mountain range, and in low-resolution models its elevation is too smoothed that it cannot efficiently block zonal winds. When we modified the elevation so that the midlatitude westerly wind is blocked and turns equatorward while descending, the anticyclonic motion in the South Pacific is strengthened, increasing the amount of low-level clouds and surface evaporation. As a result, the precipitation in the South Pacific is inhibited and the SST in the southeast Pacific is decreased.

Our sensitivity test results show that with the increasing elevation of the Andes, the east–west SST difference in tropical South Pacific increases, and the tropical Pacific SSTs decrease. Precipitation in the tropical South Pacific also decreases, and the precipitation peak is propagating westward from the central South Pacific to the western Pacific. The result also shows that the southward cross-equatorial atmospheric energy transport increases with the increasing elevation of the Andes, indicating a larger NH–SH energy imbalance, closer to the observation. Our results indicate that the control topography setting of the Andes in the CESM model with resolution of $1.9^\circ \times 2.5^\circ$ has an effective height of 2 km; the modified Andes has an effective height of 4 km.

It has been shown that the higher-resolution climate model can significantly improve the mean climate simulation. Small et al. (2014) showed that a 0.25° -resolution climate model can simulate the southeast Pacific cold tongue better than a low-resolution model. The major reason is that a better resolved orography will simulate a more accurate atmospheric flow. Adcroft (2013) considered the representation for topography as impermeable walls and porous barriers, and his results approximately simulated for the key features for lateral transport. His study provides a possible solution to fix the bias in a low-resolution global climate model.

With a better simulation of the mean tropical Pacific climate, we expect that the ENSO simulation will be improved, as well as the future projection of the global climate. However, the role of the ocean has not been tested in this work and there are still some biases in the NH that may be related to the representation of the topography in a low-resolution model. In the

future, we hope to examine the role of the ocean with a fully coupled global climate model.

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