The Effect of the Galápagos Islands on the Equatorial Pacific Cold Tongue

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ABSTRACT

A reduced-gravity ocean general circulation model of the tropical Pacific Ocean is used to determine potential improvements to the simulated equatorial Pacific cold tongue region through choices in horizontal resolution and coastline geometry—in particular, for the Galápagos Islands. Four simulations are performed, with identical climatological forcing. Results are compared between model grids with and without the Galápagos Islands, with coarse and fine resolutions. It is found that properly including the Galápagos Islands results in the obstruction of the Equatorial Undercurrent (EUC), which leads to improvements in the simulated spatial structure of the cold tongue, including a basinwide warming of up to 2°C in the east-central Pacific. The obstruction of the EUC is directly related to the improvements east of the Galápagos Islands, and for the basinwide reduction of the tropical cold bias through an equatorial dynamical adjustment. The pattern of SST warming resulting from the inclusion of the Galápagos Islands is similar to that of the known cold biases in ocean models and the current National Oceanic and Atmospheric Administration Climate Forecast System. It is thought that such an improvement would have a considerable impact on the ability of coupled ocean–atmosphere and ocean–ecosystem models to produce realistic clouds, precipitation, surface ocean bioproducivity, and carbon cycling in the tropical Pacific Ocean.

1. Introduction

The salient feature in the eastern equatorial Pacific Ocean in terms of sea surface temperature (SST) is a zonal band of minimum SST known as the cold tongue (CT). The CT is the result of coastal upwelling along the west coast of South America caused by alongshore trade winds, and upwelling along the equator caused by Ekman divergence. Seasonally, the cold tongue begins to intensify in boreal spring as coastal upwelling along the coast of Peru, prompting coupled air–sea interaction processes that effectively drive the CT westward along the equator (Mitchell and Wallace 1992; Nigam 1997). It would be difficult to overemphasize the importance of the CT in global hydrological and biogeochemical cycles, because it plays a key role in the formation of tropical cloud and precipitation patterns, the supply of nutrients for surface ocean biological productivity, and carbon cycling. The east-central tropical Pacific is the largest oceanic source of CO₂ to the atmosphere (e.g., Takahashi et al. 1999). Furthermore, large SST anomalies associated with El Niño and La Niña events are manifest as variations about the mean state of the CT, which have long been known to influence weather patterns globally (e.g., Walker and Bliss 1937; Ropelewski and Halpert 1987; Kiladis and Diaz 1989; Deser and Wallace 1990; Yulaeva and Wallace 1994).

Despite the importance of the equatorial Pacific cold tongue in global hydrological and biogeochemical cycles, most ocean general circulation models (OGCMs) and coupled atmosphere–ocean general circulation models (CGCMs) produce a CT with a cold bias, including an exaggerated westward extent (Stockdale et al. 1998; Harrison et al. 2002). The Modular Ocean Model, version 3 (MOM v.3; Pacanowski and Griffies 1999), which is the core ocean model of the present National Oceanic and Atmospheric Administration Climate Forecast System (CFS; Saha et al. 2006), exhibits a cold bias in the east-central tropical Pacific up to 1°C on the annual mean (Vecchi et al. 2005). Such a cold bias presents serious obstacles to producing realistic tropical cloud and precipitation patterns in CGCMs, including a reasonable intertropical convergence zone (ITCZ; Mechoso et al. 1995). Recent modeling studies have approached the tropical cold...
bias problem from diagnosing biases in the surface energy budget (Kiehl and Trenberth 1997), atmospheric feedbacks (Gordon et al. 2000; Sun et al. 2003), biological attenuation of shortwave radiation (Murtugudde et al. 2002; Marzeion et al. 2005), and coupled air–sea interactions (Luo et al. 2005). Despite advances in our understanding of the CT and the processes governing its mean and variability, the tropical cold bias problem remains the norm in OGCMs and CGCMs.

The existence of the Galápagos Archipelago on the equator near 90°W, made famous after the nineteenth-century expeditions of British naturalist Charles Darwin, presents the potential for topographic interaction with the equatorial current system and other processes related to the CT. Current operational ocean analysis and prediction systems [e.g., the Global Ocean Data Assimilation System (GODAS), the oceanic component of the NOAA CFS] do not include the Galápagos Islands (D. Behringer 2006, personal communication). Early observational and theoretical studies examined the impact of the Galápagos Islands on the structure of the Equatorial Undercurrent (EUC) in the eastern Pacific (e.g., Stevenson and Taft 1971; Christensen 1971; White 1973; Lukas 1986). All observational analyses indicate that the core of the EUC east of the Galápagos Islands is at least reduced. Ship-based velocity measurements [e.g., World Ocean Circulation Experiment (WOCE) and N. B. Palmer sections] disagree on whether EUC remnants flow north or south of the islands (W. S. Kessler 2006, personal communication). Eden and Timmerman (2004, hereinafter ET04) examined topographic effects on the EUC, South Equatorial Current (SEC), and tropical instability waves (TIWs) by comparing output from an OGCM including the Galápagos Islands with that in which the islands’ subsurface topography above 2000-m depth was removed. The horizontal resolution of the ET04 experiments was \( \frac{1}{4}° \times \frac{1}{4}° \). Their analysis of SST focused on TIWs and changes in total eddy kinetic energy.

What is needed is a direct assessment, including validation against observations, of potential improvements to the tropical cold bias problem through choices in horizontal resolution and grid geometry in an OGCM that does not prescribe air–sea heat fluxes as a Newtonian damping condition. The present study is such an assessment, employing a well-tested OGCM, with particular focus on the Galápagos Islands. The ability to make some comparisons with the results of ET04 and others is an asset to the authors in interpreting results. A description of the model, datasets used, and experimental philosophy are provided in the following section. The results are presented in section 3, and a summary is found in section 4.

2. Experimental philosophy

The objective is to quantify and understand the relative improvements to the simulated mean and climatology of the equatorial Pacific cold tongue region in an OGCM arising from increased horizontal resolution and the inclusion of the Galápagos Islands. The model chosen for the present study is the Gent and Cane (1989) reduced-gravity primitive equation model of the tropical oceans with a hybrid vertical mixing scheme (Chen et al. 1994). The Chen et al. (1994) scheme accounts for mixed layer entrainment–detrainment, shear flow instability, and free convection in the thermocline by combining the physics of the Kraus and Turner (1967) mixed layer model with the Price et al. (1986) dynamical instability model. Surface fluxes are calculated interactively by coupling the OGCM to a thermodynamic atmospheric mixed layer (Murtugudde et al. 1996), thus allowing for feedbacks between SST and surface fluxes. Originally developed for modeling the equatorial Pacific Ocean, the Gent–Cane OGCM is structured vertically on sigma coordinates and includes a mixed layer plus 19 subsurface layers. The meridional boundaries of the model grid are 40°N–40°S, along which a sponge layer is used at the meridional open boundaries (see Chen et al. 1994), and the Indonesian Throughflow is closed off. The Gent–Cane OGCM has previously been used to assess the mean seasonal heat budget of the CT and accurately reproduced many aspects of the observed climatology in the eastern equatorial Pacific Ocean (Kessler et al. 1998; Chen et al. 1994). The model was spun up from the climatology of Levitus and Boyer (1994) for 60 yr. Four 1-yr integrations under identical climatological forcing were carried out, beginning with a properly spun up state for each case. Climatological forcing consisted of 2° \( \times \) 2° European Centre for Medium-Range Weather Forecasts (ECMWF) operational analysis surface wind stress, with a base period of 1985–2003 (Bengtsson et al. 1982); Xie and Arkin (1996) precipitation, with a base period of 1979–2003; International Satellite Cloud Climatology Project (ISCCP) cloud cover, with a base period of 1983–94 (Rossow and Schiffer 1991), and Earth Radiation Budget Experiment (ERBE) shortwave radiation, with a base period of 1984–90 (Barkstrom 1984).

A total of four cases are examined. The first case (coarse) was run on a grid with uniform zonal resolution (\( \frac{1}{4}° \)), and meridional resolution stretching from \( \frac{1}{8}° \) at the equator to \( \frac{1}{16}° \) at the meridional boundaries, without the Galápagos Islands. The second case (coarse+G) was run on the same grid as that of the coarse case, but included the Galápagos Islands. The third case (fine) was run on a grid with increased reso-
olution in the eastern tropical Pacific, with zonal stretching from $\frac{1}{4}^\circ$ in the east to $1^\circ$ at the western boundary, and meridional stretching from $\frac{3}{4}^\circ$ at the equator to $1^\circ$ at the meridional boundaries, without the Galápagos Islands. The final case ($fine+G$) was run on the same grid as that of the fine case, but included the Galápagos Islands. For computational efficiency, the meridional limits of the grids used in the fine and fine+$G$ cases were reduced to $30^\circ N$–$20^\circ S$. The fine and fine+$G$ experiments are of similar horizontal resolution as those of ET04.

Implementation of the Galápagos Islands is subject to the constraints of each model setup. The Galápagos Archipelago is a group of 19 islands roughly 1000 km off the coast of Ecuador. Most of the islands making up the Galápagos Archipelago are miniscule by comparison, except for the largest, Isla Isabela. Accounting for 73% of the total land area of the Galápagos Islands, Isla Isabela occupies 5825 km$^2$, which makes it larger than the state of Delaware. In light of this inequity, the Galápagos Islands were implemented for the fine+$G$ case as land points on the model grid approximating the size and shape of Isla Isabela, and similarly for the coarse+$G$ case, but with less detail (Fig. 1). The model setup used in ET04 included the partial cell topography improvements of Pacanowski and Gnanadesikan (1998); however, there is little if any variation with depth of the topography of the Galápagos Islands in ET04. The case in the Gent–Cane OGCM is similar, in which all topography is constant in the vertical.

Observational datasets used for comparison with model output include the $2^\circ \times 2^\circ$ Reynolds and Smith (1994) optimal interpolation (OI) SST (often referred to hereinafter as “Reynolds”), with a base period of 1982–2005; the $\frac{1}{4}^\circ \times \frac{1}{4}^\circ$ Tropical Rainfall Measuring Mission (TRMM; Kummerow et al. 2000) Microwave Imager (TMI) SST, with a base period of 1998–2005, and Tropical Atmosphere–Ocean (TAO) subsurface temperature and currents (Hayes et al. 1991), with a base period of 1980–2005. TMI is needed to show the detailed spatial structure that is smoothed out in Reynolds OI, but because of TMI’s La Niña bias, Reynolds OI will also be used to show the real seasonal cycle.

3. Results

a. Comparison of simulations and observations

The annual mean SST fields from the four simulations and TMI are shown in Fig. 2. There appears to be very little difference when either (a) adding the Galápagos Islands under coarse resolution (i.e., coarse versus coarse+$G$), or (b) increasing the horizontal resolution without the Galápagos Islands (i.e., coarse versus fine). However, adding the Galápagos Islands under fine resolution (i.e., fine versus fine+$G$) has the effect of reducing the westward extent of the CT. In fact, the westward extent of the CT appears to be underestimated when compared with the TMI climatology (with a base period of 1998–2005), while the
structure closer to the Galápagos Islands and the coast of South America is in better agreement than any other case.

To analyze the simulated mean seasonal cycle of SST in the CT region, an index was constructed as the area-averaged SST from 170°W to the coast of South America, from 1°N to 1°S (indicated in Fig. 2, bottom panel). The resulting seasonal cycles from all four cases and Reynolds OI are displayed in Fig. 3. Reynolds OI is used here because its longer temporal record should make it more representative of climatology while spatial comparisons are not needed. Under coarse resolution, the inclusion of the Galápagos Islands made little difference; the coarse and coarse+G cases exhibit a similar cold bias all year, which is more pronounced during the boreal spring and autumn seasons and consistent with the global models (Stockdale et al. 1998). Under fine resolution, however, the Galápagos Islands have a warming effect on CT SST anywhere from 0.8° to 1.5°C. Throughout the year, CT SST in the fine+G case is closer to observations than in that in the fine case; the root-mean-square error (RMSE) with respect to Reynolds OI is reduced from 1.0° to 0.3°C (70% reduction). The longitudinal distribution of the biases in each case is shown in Fig. 4. In the east-central part of the basin, there is little difference between the coarse and coarse+G cases. East of the islands, however, the inclusion of the Galápagos Islands warms the SST anywhere from 0.25°–1°C. In the fine-resolution experiments, the effect of the Galápagos Islands is to warm SST everywhere except for a very narrow (~2° longitude wide) region immediately west of the islands. In the central region, a relatively strong warm bias (~1°C) actually develops in the fine+G case that was only moderate (~0.5°C) in the fine case, but everywhere east of ~140°W the cold bias is much reduced. For reference, the difference between TMI and Reynolds SST is also provided. TMI is clearly colder than Reynolds throughout the east-central equatorial Pacific, but it also resolves the cold (warm) signal immediately west (east) of the Galápagos Islands.

Diagnosing changes in seasonal variability between
The simulations is not the primary focus of the present study, but it is worthwhile to note that the seasonal variability of SST is also improved in the east-central region by including the Galápagos Islands under fine resolution (Fig. 5). Under coarse resolution, there is some reduction of the westward extent of high seasonal variability (standard deviations of 1°C or more are shaded). However, the region of high seasonal variability in the fine+G case is characterized as broader and shorter, similar to Reynolds, rather than narrow and long as in the other three experiments.

The improvements in the cold tongue region by increased resolution and the Galápagos Islands are also evident upon comparing the spatial structure of seasonal mean SST fields. Based on observed climatology and output from all four simulated cases, the CT is least developed in terms of coastal and equatorial SST in boreal spring [March–May (MAM)] and most developed in boreal autumn [September–November (SON)]. Figure 6 (Fig. 7) is a comparison of the mean MAM (SON) SST in the east-central equatorial Pacific Ocean between all four simulated cases and TMI observations. In both seasons, the combined effects of resolution and the Galápagos Islands are similar. Under coarse resolution, the improvements are local—slight warming immediately east and cooling immediately west of the Galápagos Islands—while the coastal upwelling signal and westward extent of the CT are relatively unchanged. Under fine resolution, three improvements are found: the spatial structure of the CT immediately east and west of the Galápagos Islands (which clearly indicates the presence of an island), the structure of the coastal upwelling signal (the coldest water is tightly confined to the coast of South America), and a reduction in the westward extent of the CT. Improvements from resolution alone are relatively minor. However, improvements resulting from the Galápagos Islands are only impressive under fine resolution.

Different processes may be responsible for SST changes in the east-central part of the basin and in the region east of the Galápagos Islands. The CT index used thus far, spanning 90° longitude, is too broad for interpreting the response east and west of the Galápagos Islands. Therefore, two small (2° × 2°) boxes centered on the equator, one east of the Galápagos Islands (85°W) and one to the west (110°W), were chosen for comparison of simulated mean seasonal cycles of SST and Reynolds OI (indicated in Fig. 7, bottom panel). Under coarse resolution (Fig. 8), there is little change in SST west of the islands, while at 85°W, the inclusion of the islands increases SST up to 1°C, bringing the simulated mean seasonal cycle in closer agreement with ob-

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**Fig. 3.** Mean seasonal cycle (monthly means) of SST (°C) in the cold tongue index region (1°S–1°N, 170°–80°W) for the four simulated cases and the Reynolds SST climatology. Lines are as indicated on figure key.

**Fig. 4.** Annual mean meridionally averaged (1°N–1°S) (left) SST bias (°C, with respect to Reynolds) and (right) TMI across the index area shown in Fig. 2 of four simulated cases. Lines are as indicated on figure keys.
servations. This warming to the east of Galapagos is caused by a weakening of the thermocline and reduced entrainment mixing, and is similar to the changes in the fine+G simulation. Under fine resolution, the simulated mean SST east and west of the Galápagos Islands is considerably improved (Fig. 9). In addition to the surface, temperature differences in excess of 2°C between the fine and fine+G cases are found at thermocline depth (Fig. 10). Along the equator, the largest positive temperature difference between the fine+G and fine cases is found at approximately 75 m. Likewise, along 110°W the largest difference is found at 75 m, but differences of over 1.5°C extend to depths of 250 m or more, which is consistent with the changes in EUC and SEC noted in the following section.

In the equatorial Pacific Ocean, zonal transport is dominated by the eastward EUC and the westward SEC. Changes to the EUC, a most vital component to the dynamical balance of the equatorial Pacific Ocean, are also noted under fine resolution (Figs. 11–14). Presented in Fig. 11 is a comparison of vertical profiles of zonal velocity at 110°W (about 20° longitude west of the Galápagos Islands) between the four simulated cases and available TAO depths. The effect of increased model resolution on zonal velocity at 110°W is to lower the core of the EUC by 10–20 m, and increase its speed by over 30 cm s\(^{-1}\). The faster EUC in the fine-resolution simulations could be partly attributed to stronger meridional gradients of zonal velocity, which has been shown to result in the eastward advection of cyclonic relative vorticity and thus a stronger EUC than would be expected from the zonal pressure gradient alone (Kessler et al. 2003). Furthermore, near the core of the EUC (100 m), the inclusion of the Galápagos Islands under fine resolution (fine+G) brings the simulated zonal velocity to within 2.4 cm s\(^{-1}\) of the TAO observations. The profile of zonal velocity at 110°W from the climatology of the existing version of the National Centers for Environmental Prediction (NCEP) Global Ocean Data Assimilation System (GODAS) is similar to the coarse-resolution profiles in Fig. 11 (Behringer and Xue 2004; cf. Fig. 7). The SST improvements immediately east and west of the Galápagos Islands, as will be discussed in the following section, are related to the obstruction of the EUC by the Galápagos Islands in the fine+G case. We will also show that the improvements to the broader tropical cold bias problem are due to the obstruction of the EUC as well, but through a basinwide dynamical and heat flux adjustment that effectively produces an
SST warming across the east-central tropical Pacific Ocean.

b. Influence of the Galápagos Islands on the EUC

A case is presented here in which the simulated improvements to the tropical cold bias problem begin with the obstruction of the EUC by the Galápagos Islands at 92°W. From the comparison of vertical sections of temperature and zonal velocity along the equator (Fig. 12), it can be seen that zonal velocity in the EUC approaches zero at the western edge of the Galápagos Islands in the fine+G case, while in the fine case, the EUC is allowed to continue unobstructed until reaching the continental shelf of South America. A similar, although much more gradual, reduction in the EUC approaching the Galápagos Islands is evident under coarse resolution. The effect of obstructing the EUC on the upper-ocean thermal structure east of the Galápagos Islands is a deeper thermocline; the 20°C isotherm is over 10 m deeper along the coast of South America in the fine+G than in the fine case. Because the mean thermocline in the eastern equatorial Pacific is very shallow to begin with, relatively small changes in the depth of the thermocline can have a large impact on SST. In comparing the subsurface thermal structure west of the Galápagos Islands, it is seen that a given isotherm outcrops at least 10° longitude farther to the east in the fine+G than the fine case, appreciably reducing the cold bias that is present without the Galápagos Islands.

From a zonal depth section (e.g., Fig. 12) one cannot determine whether the EUC terminates, is deflected, or splits around the Galápagos Islands. ET04 compared latitude–longitude plots of zonal velocity at 80-m depth in the eastern equatorial Pacific Ocean between simulations with and without the Galápagos Islands (cf. Fig. 3). In the model simulation of ET04 that included the Galápagos Islands, the EUC is deflected northward around the island where it returns to the equator and continues on its original path. For that reason, ET04 found only localized changes to the SST field (i.e., warming immediately east of and cooling immediately west of the Galápagos Islands). Figure 13 shows the analogous zonal velocity depiction for the present
study, comparing the fine and fine+G cases. Clearly evident in the fine+G case is an obstruction or, at least zonally, a near termination of the EUC at the western boundary of the Galápagos Islands. This near termination of the zonal component of the EUC is further evident in Fig. 14, which is a series of meridional sections of zonal velocity 2° west of, through, and 2° east of the Galápagos Islands. Along each of the meridians shown for the fine case (top), there is little change in the strength or location of the EUC. In the fine+G (bottom), as in fine, case the EUC has a core of roughly 70 cm s\(^{-1}\) at 93°W; but, upon meeting Isla Isabela, a weak (10–20 cm s\(^{-1}\)) branch is directed south, and another is directed north, of the barrier. At 89°W, all that remains of zonal velocity is a slightly deeper 10 cm s\(^{-1}\) lobe to the north of the original core.

It is instructive to understand the fate of the EUC following its encounter with the Galápagos Islands. In terms of volume transport (integrated from 4°S to 4°N and 0–300 m), eastward volume transport \(T_x\) is reduced by 62% between 95° and 85°W. To estimate how much of that reduction in \(T_x\) is being compensated by poleward volume transport \(T_y\) versus vertical transport \(T_z\), \(T_y\) was calculated by integrating meridional velocity from 95° to 85°W and 0–300 m at 4°N and 4°S. Poleward transport, or the sum of northward \(T_y\) through the 4°N plane and southward \(T_y\) through the 4°S plane, only accounts for 15% of the amount by which eastward \(T_x\) was reduced across the Galápagos Islands. This implies that \(T_z\), taken as the residual, accounts for the remaining 85% of the mass balance. Furthermore, 86% of the poleward transport is accounted for by the southward \(T_y\), which becomes available to feed the Peru–Chile Undercurrent.

As a demonstration of the improved modeling of subsurface flow resulting from the inclusion of the Galápagos Islands, a comparison is made with the seminal work of Lukas (1986). Lukas (1986) observed remnants of the EUC flowing southeastward from the Galápagos Islands to converge with the Southern Subsurface Countercurrent and form the Peru–Chile Undercurrent around 100-m depth. The location of the

Fig. 7. As in Fig. 2, but for September–November; the heavy contour is now 26°C, and the SST colder than 22°C is now shaded. The yellow boxes indicate the index regions used in Figs. 8–9.
Peru–Chile Undercurrent was then corroborated with inferences from dynamic topography. Based on a comparison between current vectors and speed at 100-m depth between the fine and fine +G cases (Fig. 15), the joining of the EUC remnants southeast of the Galápagos Islands with the Southern Subsurface Countercurrent and the formation of the Peru–Chile Undercurrent is evident in the fine +G case, confirming the important role of the interaction between the Galápagos Islands and the EUC in its formation. Both cases exhibit a narrow corridor of southward flow along the coast, but the connection with the EUC, as in the analyses of Lukas (1986; cf. Fig. 17), is only evident in the fine +G case.

c. Mechanisms for reduction of tropical SST bias

The more remarkable improvement to the simulated climatology of the tropical Pacific Ocean owing to the Galápagos Islands is not necessarily the fact that the EUC is blocked well short of the South American coast, but rather the positive mean SST difference throughout the east-central tropical Pacific. The existence of the Galápagos Islands has a warming effect on a broad region from 10°S to 10°N, up to 2°C (Fig. 16a).

Also shown in Fig. 16a is the difference in the ocean–atmosphere net heat flux (NHF) between the fine +G minus fine cases. Differences of 10–20 W m⁻² are found across the east-central tropical Pacific. This points to the important role of the feedback between SST and ocean–atmosphere heat flux, and could have an appreciable impact on the thermodynamics of the lower troposphere. A straightforward heat budget analysis was performed on the two small boxes used for the SST mean seasonal cycle comparison in section 3. Each term was contrasted between the fine and fine +G cases to determine what processes are responsible for the warming previously shown and discussed. In the 85°W box east of the Galápagos Islands, vertical entrainment mixing is reduced; the difference in the annual mean contribution to the mixed layer heat budget by entrainment mixing is 15.8 W m⁻² (Fig. 16b). Meridional and zonal thermal advection also contribute to the warming east of the Galápagos Islands (Figs. 16c,d). West of the Galápagos Islands, the difference in contribution to the mixed layer heat budget by entrainment mixing remains broad and positive (+13.4 W m⁻² in the 110°W box), while differences resulting from horizontal thermal advection represent the signal of westward-propagating waves and average to zero over time and space.

Given that reduced entrainment mixing is the domi-
nant term responsible for the positive SST difference between the fine and fine+G cases, the remainder of this section is aimed at establishing why entrainment mixing would be reduced in the fine+G case. Two important factors that determine the effectiveness of entrainment mixing are the vertical gradient of temperature, particularly at the base of the mixed layer, and the depth of the thermocline. Across the entire east-central tropical Pacific, the vertical temperature gradient at the mixed layer depth is more diffuse in the fine+G than the fine case. For example, using once again 110°W and averaging from 2°N to 2°S, the vertical temperature gradient at mixed layer depth is more diffuse in the fine+G (−0.10°C m⁻¹) than the fine (−0.13°C m⁻¹) case. While these differences appear small, note that these are the remaining differences as a net result of the large-scale dynamic and thermodynamic adjustments, including shear mixing and entrainment under prescribed forcing.

For a given zonal wind stress and ocean geometry, there will be an equilibrium mass and energy balance. In the Pacific Ocean, the dynamical features maintaining this balance are the easterly trade winds, the westward SEC, the southward interior Sverdrup flow, a thermocline that slopes up to the east, a sea surface that slopes up to the west (mirroring the thermocline), a westward zonal pressure gradient, and the EUC flowing eastward along the thermocline. After sufficient spinup time, such is the case in a model. However, should one component of this balance become disrupted, a new, adjusted dynamical balance must follow. An attempt at illustrating this adjustment using output

**Fig. 10.** Zonal depth sections along the equator of September–November mean temperature (°C) difference between the fine+G minus fine cases, and the corresponding meridional section along 110°W.

![Zonal depth sections](image)

**Fig. 11.** September–November mean vertical profiles of zonal velocity (m s⁻¹) at 0°N, 110°W for the four simulated cases. Yellow bands represent corresponding TAO zonal currents at 15, 50, 75, 100, and 150 m. The width of the yellow bands represents the extent to which the values depend on the climatological base period used (i.e., 1980–89, 1990–2005, or 1980–2005).
from the fine and fine+G simulations is presented in Fig. 17. The disruption in this case is the obstruction of the EUC at the Galápagos Islands, which is evident in the difference in EUC core velocities around 92°W (“EUC core velocity” is the maximum positive zonal velocity found between 5°S and 5°N at any depth). In response to this disruption, equatorial sea level rises across the basin. However, because EUC transport is also reduced basinwide, the change in sea level is non-uniform; the sea level rise in the west is less than that in the east, resulting in a more gradual sea surface slope. Also evident in Fig. 17 is the fact that the SEC in the east-central region is approximately 10 cm s\(^{-1}\) slower in the fine+G case, or that there is a roughly 25% reduction in SEC velocity between 160° and 90°W (“SEC velocity” is the surface zonal velocity averaged from 5°S to 5°N). Corresponding changes in the subsurface thermal structure are also evident; the base of the mixed layer is deeper between 170° and 130°W, and the 20°C isotherm is deeper throughout the east-central re-

Fig. 12. Vertical sections of September–November mean temperature (°C; shaded) and zonal current (m s\(^{-1}\); contoured) along the equator in the upper 300 m of the Pacific Ocean for the simulated (a) coarse, (b) coarse+G, (c) fine, and (d) fine+G cases. Temperature warmer than 20°C is shaded every 2°C, zonal currents are contoured every 10 cm s\(^{-1}\), and the heavy contour is 50 cm s\(^{-1}\).

Fig. 13. September–November mean zonal currents (m s\(^{-1}\)) at 80-m depth in the eastern equatorial Pacific Ocean for the simulated fine and fine+G cases. The contour interval (CI) is 20 cm s\(^{-1}\), the zero contour is omitted, and negative values are dashed.
region where positive SST differences were found. When compared with the climatology of the September–November mean 20°C isotherm depth from TAO measurements, closer agreement is found with the fine+G case at all available longitudes (Fig. 17). East of 180°W, the difference in the 20°C isotherm depth represents an over 50% reduction in RMSE with respect to TAO measurements. The new, adjusted equilibrium state of the equatorial Pacific in fine+G caused by the obstruction of the EUC at the Galápagos Islands, leading to reduced EUC–SEC shear, reduced westward advection of cold SST by the SEC, and a deeper thermocline lends itself to the positive SST differences found in nearly the same location as that of known SST cold biases. None of the aforementioned differences highlighting the zonal dynamical adjustment were discernable between the coarse and coarse+G cases, which is undoubtedly because the EUC did not extend far enough to the east for the Galápagos Islands to actually obstruct it (e.g., Fig. 12).

Important differences in meridional circulation were also found between the fine and fine+G cases, which are not unrelated to the zonal adjustment. The meridional circulation in the equatorial Pacific Ocean is characterized by import at thermocline depth and export at the surface. Given such a configuration, mass continuity mandates upwelling and entrainment of cold water at the equator. This circulation is present in both fine and fine+G, but is considerably slower in the fine+G case. At 4°N, averaged from 180° to 100°W, the maximum southward velocity (import) is reduced by 15%, and the maximum northward velocity (export) is reduced by 65% from fine to fine+G. At 4°S, there is little change in maximum northward velocity (import), but maximum southward velocity (export) is reduced by 27%. Translating to an overall reduction in the rate of meridional equatorial overturning of roughly 25%, an increase in SST follows. As would be expected from a more diffuse thermocline, the aforementioned zonal adjustment, and reduced meridional circulation, entrainment mixing is considerably weaker and less effective, resulting in a pattern of reduced cooling that approximates the pattern of SST warming (Fig. 16b). The reader is reminded that the wind stress forcing was
identical in all cases, thus any changes in circulation are independent of any wind-driven processes such as Ekman divergence. One could say that many aspects of the ocean’s response to the surface winds that act to produce zonal inhomogeneities (e.g., 20°C isotherm depth, sea level, SST, etc.) are lessened without requiring a change in zonal wind stress. Clearly, a warmer CT and a reduced zonal SST gradient will have local and remote wind responses in a coupled model and as pointed out by Schneider and Zhu (1998); they will further alleviate the cold bias in the CT.

4. Summary and discussion

The objective of the present study was to assess the potential improvements to the equatorial Pacific cold tongue region resulting from a higher spatial resolution and the inclusion of the Galápagos Islands. It was found that the Galápagos Islands obstruct the EUC, which prompts a basinwide adjustment in the equatorial mass and energy balance. The result of the dynamical adjustment is a deeper and more diffuse thermocline, as well as reduced meridional adjustment in the equatorial mass and energy balance. The result of the dynamical adjustment is a deeper and more diffuse thermocline, as well as reduced meridional overturning at the equator. All of these results lead to reduced entrainment mixing, and therefore warmer SST. The feedback between SST and oceanic–atmospheric heat flux also contributes to the warming.

In many respects, simply increasing the resolution without including the Galápagos Islands did not result in improvements, but instead exacerbated the cold bias and produced an EUC that was too strong east of where the islands should be. On the other hand, differences resulting from the Galápagos Islands without increasing the horizontal resolution were negligible. Only when the Galápagos Islands were given proper treatment with sufficient horizontal resolution did a more realistic depiction of the CT and a reduction in the tropical cold bias problem emerge. In other words, the horizontal resolution must be fine enough to produce an EUC that extends far enough eastward, but the Galápagos Islands must be there to obstruct it.

In the experiments of ET04, the EUC deflects north and promptly returns to its original path without extensive loss of zonal velocity. The reasons for such drastically different EUC responses to the Galápagos Islands between the present study and that of ET04 are straightforward. Although there are differences in the OGCMs used in ET04 and the present study (horizontal resolution is not one of them), it is thought that the differences in the results arise from the fact that the Galápagos Islands in ET04 were too small (a meridional extent about $\frac{3}{4}$°) and are entirely south of the equator (by about $\frac{1}{4}$°). In the present study, the Galápagos Islands were implemented into our OGCM.
with realistic meridional extent and location. Furthermore, the EUC in the ET04 simulation that includes the Galápagos Islands is centered directly on the equator, rather than $1^\circ S$ where it is known to be. For these reasons, the EUC simply “glances” the Galápagos Islands in ET04, which at least partially explains why ET04 did not report a broader tropical SST warming. Differences in the surface heat flux formulation also contribute to the differences in the extent of air–sea interactions between the two models (see Murtugudde et al. 1996).

The spatial pattern of the SST difference fine+G minus fine is similar to the known biases in the MOM3 OGCM and the NCEP GODAS. It is thought that such an improvement, including the reduced atmosphere–ocean NHF, would have a considerable impact on the ability of the coupled ocean–atmosphere and ocean–ecosystem models to produce realistic clouds, precipitation, biological activity, and carbon cycling in the tropical Pacific Ocean. It is clear that the effects of the Galápagos Islands on the Pacific Ocean, beginning with equatorial currents, cannot be ignored on the annual mean. Future work should be aimed at exploiting similar effects on interannual variability and the implications for the predictability of the coupled ocean–atmosphere system. In particular, the potential benefits to the predictive skill of, for example, future iterations of the NCEP GODAS and CFS by implementing the Galápagos Islands at a sufficient resolution should be given serious consideration.

Fig. 16. September–November mean SST difference (°C; shaded): (a) net ocean–atmosphere heat flux difference (contour interval 10 W m$^{-2}$), (b) difference in contribution to the mixed layer heat budget by entrainment mixing (W m$^{-2}$), (c) zonal thermal advection, and (d) meridional thermal advection for fine+G minus fine cases.
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