Progress in the 3-D circulation of the eastern equatorial Pacific in a climate ocean model

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Abstract

The Pacific Equatorial Undercurrent (EUC) displays important seasonal, interannual and decadal variability. The correct representation of the EUC in ocean general circulation models (OGCM) is of primary importance for an adequate simulation of the Tropical Pacific Ocean, and for the ENSO (El Nino Southern Oscillation) cycle. This study aims at better understanding the EUC termination and its interactions with the three-dimensional circulation in the eastern Pacific in a climate-type OGCM. It also aims at improving the realism of the EUC in this region where some deficiencies were previously noticed. Several sensitivity experiments have been performed from 1993 to 2000, and the model results are compared to TAO observations. The sensitivity to vertical and lateral eddy mixing coefficients is explored, and we find that the 3-D equatorial circulation is highly sensitive to a reduction of the background vertical eddy coefficients in the surface layers. This reduction is only possible because of the change in the advection scheme and leads to a great improvement in the annual mean and the seasonal cycle of the modelled EUC, increasing both the EUC and the tropical cells transports. However, it appears that in the eastern Pacific, the increase in EUC transport is largely independent of the increase in the 3-D circulation. This is also true for the seasonal cycle transport. Reducing the isopycnal mixing coefficients further improves the mean EUC speed, but does not modify the associated 3-D circulation. Adding the Galapagos Archipelago to the model bathymetry only induces very local changes in the equatorial 3-D circulation. However, it leads to important changes in the pathways of EUC waters that reach the South American coast.

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1. Introduction

The Equatorial Undercurrent (EUC) is a major feature of the tropical Pacific circulation. This strong subsurface eastward flow is tightly confined to the equator and lies within the equatorial thermocline from north
of New Guinea to the coast of South America. It starts at about 200 m depth around 140°E and shoals towards the east, following the trend of the thermocline. Its cold and nutrient-rich waters are directly feeding the equatorial upwelling in the eastern Pacific, giving rise to the cold tongue and influencing the complex heat budget in this region (Bryden and Brady, 1985; Kessler et al., 1998). The EUC waters are entrained into the surface layers through upwelling and mixing and modify the sea surface temperature (SST). Hence, they may have a strong influence on the coupling between the ocean and the atmosphere (Cronin and Kessler, 2002) and on ENSO (El Niño Southern Oscillation), the dominant mode of variability in the tropics.

In the far eastern Equatorial Pacific, the termination of the EUC and its interactions with the three-dimensional circulation are complex and not well understood (Kessler, 2006). The presence of the Galapagos Islands in the equatorial band also complicates the picture (Brennemann, 1999). The cold waters upwelled from the EUC diverge poleward through Ekman transports and feed the SEC. Some of the diverging surface waters downwell at 4°–5°N (4°–5°S) because of negative (positive) wind stress curl respectively, and converge back towards the EUC in the pycnocline, forming the Tropical Cells (TC) and partly compensating for the Ekman divergence (McCreary and Lu, 1994; Johnson, 2001; Sloyan et al., 2003). They participate in the mixing and the recirculation of the upper part of the EUC. Entrainment of EUC waters into the upper layers also contributes to its deceleration and its termination (Pedlosky, 1987; Pedlosky, 1988). At deeper levels, the EUC waters diverge northward or southward as they approach the South American coast, notably feeding the Peru–Chili current system (Lukas, 1986). However, the exact pathways of the EUC waters and their variability are poorly known.

Strong seasonal and interannual variations in the transport, the temperature and the position of the EUC do exist (Johnson et al., 2002). During boreal spring, the EUC strengthens and surfaces (Yu and McPhaden, 1999; Yu et al., 1997). In the central Pacific, it has been shown that the TC recirculating waters play an important role in the seasonal variability of the EUC transport (Blanke and Raynaud, 1997). Whether such a link exists in the eastern Pacific has not yet been determined. During El Niño events, the EUC weakens and may even disappear (e.g. Kessler and McPhaden, 1995; Johnson et al., 2002). In addition, it has been suggested that the large variations in transport and temperature of the EUC in 1998 had an influence on the rapid turn from El Niño to La Niña conditions in June (Izumo et al., 2002). The EUC is also part of the subtropical meridional overturning cells (e.g. McCreary and Lu, 1994) by which subducted subtropical waters can influence the equatorial temperatures at decadal timescales (Gu and Philander, 1997). At all of these timescales, the EUC variability certainly modifies the Peru current system, which is one of the most productive upwelling region in the world.

An adequate simulation of the EUC, of its 3-D circulation and of its termination is thus essential in ocean general circulation models (OGCM). It is also crucial for the ability of coupled general circulation models to simulate realistic ENSO oscillations, and decadal variability. In the global configuration ORCA2 of the OPA OGCM (Madec et al., 1998), which is used for a wide range of oceanographic and climatic studies, notably used for IPCC (Intergovernmental Panel on Climate Change) and paleoclimate simulations (Braconnot et al., 1999), the simulated EUC is very realistic (e.g. Vialard et al., 2001; Lengaigne et al., 2003). However, in the Lengaigne et al. (2003) simulations, the magnitude of the mean simulated EUC in the eastern Pacific is weaker than observed by more than 20 cm s\(^{-1}\) at 110°W. At the surface, the mean westward South Equatorial Current (SEC) is too strong and too deep compared to the observations. This bias can be observed in other OGCMs (e.g. Large et al., 2001). In this region where the thermocline is very shallow, an accurate simulation of the mean equatorial currents is of particular importance for an accurate simulation of the SST.

What are the physical causes of this deficiency? How can we improve the representation of the equatorial currents in this region? Part of the weakness of the OGCM simulations lies with the inaccurate representation of subgrid-scale oceanic processes, and in the parameterisations of momentum and tracer turbulent mixing. Yet, mixing is of primary importance for the EUC simulation, and it has been shown that both vertical and lateral mixing are important terms of the EUC zonal momentum equation (Bryden and Brady, 1989; Qiao and Weisberg, 1997; Wacongne, 1989; Maes et al., 1997). Maes et al. (1997) conducted sensitivity experiments by drastically decreasing the horizontal mixing in the z-coordinate OPA model. They concluded that in the eastern Pacific, the EUC is mainly decelerated by horizontal advection and vertical diffusion in its upper part, and by the horizontal diffusion in its lower part. Megann and New (2001) also investigated the effect of
reducing the viscosity on the equatorial circulation in an isopycnal model. They showed that reducing the viscosity leads to a stronger and narrower EUC simulation. Lateral and vertical momentum mixing thus act as a brake for the equatorial circulation, and the weakness of the simulated EUC may be due to an unrealistically strong vertical and lateral turbulent mixing in this region. Another possible weakness of the modelled currents in the Eastern Pacific could be due to the bathymetry. For example, in many models, the Galapagos Islands do not reach the surface. It is suspected that the presence of the Galapagos Islands in the simulations may influence the equatorial circulation in the Eastern Pacific (Brennand, 1999), and may reduce the mean surface currents in the eastern Pacific, as suggested by the study of Eden and Timmermann (2004).

The goal of this study is to improve the EUC simulation in the global configuration ORCA2 (based on a 2° Mercator mesh) of the OPA OGCM, and to better understand the sensitivity of the tropical cells and of the EUC termination to different parameterisations. We perform sensitivity experiments whereby we decrease both the vertical and lateral eddy viscosity and diffusivity coefficients used to calculate isopycnal mixing in the tropics. We also perform a simulation that includes the Galapagos Islands, and explore the impact of this on the currents and temperatures in the Equatorial Pacific Ocean. The paper is organized as follows. Section 2 describes the model and the different sensitivity experiments. Section 3 compares the mean currents of the experiments, explores the physical mechanisms involved in their differences and studies the sensitivity of the mean tropical cells and of the EUC termination to the different parameterisations. It also compares the mean currents of the experiments to observations. Section 4 examines the EUC and its link to the associated seasonal cycles of the 3-D circulation. In Section 5, the impact of the different parameterisations on the temperature fields is studied. Section 6 provides a discussion and concludes on our results.

2. Model description and experiments

The OGCM used in this study is the OPA model (Madec et al., 1998) in its global configuration ORCA2. The horizontal mesh is based on a 2° by 2° Mercator grid, and following Murray (1996), two numerical inland poles have been introduced in order to remove the North Pole singularity from the computational domain. The departure from the Mercator grid starts at 20°N, and is constructed using a series of embedded ellipses using the semi-analytical method of Madec and Imbard (1996). In addition, the meridional resolution is increased in the tropics, up to 0.5° at the equator. There are 31 vertical levels from the surface to the bottom with a resolution of 10 m in the upper 150 m, decreasing to 500 m in the deep ocean. The model uses a free surface formulation (Roulet and Madec, 2000) and computes the density from potential temperature, salinity and pressure using the Jackett and McDougall (1995) equation of state.

Lateral tracer and momentum mixing is performed along isopycnals (e.g. Redi, 1982), as it has been shown that this parameterisation better simulates the vertical structure of the observed eddy effects in the tropics (Lengaigne et al., 2003) and significantly improves the structure of the simulated EUC in the OPA OGCM. As shown in their paper, the rotation of the momentum tensor reduces the lateral mixing in the upper part of the equatorial thermocline, and intensifies the upper part of the EUC. In addition, a strong horizontal background viscosity is added poleward of 10°, smoothly increasing from 0 at 10° to 40,000 m² s⁻¹ poleward of 20°. The vertical eddy viscosity and diffusivity coefficients are of primary importance for our study. They are computed from a 1.5 turbulent closure scheme based on a prognostic equation for the turbulent kinetic energy (Blanke and Delecluse, 1993) and a diagnostic equation for the dissipation and mixing turbulent length scales. In locations with a statically unstable stratification, convective mixing is simulated by assigning a value of 100 m² s⁻¹ to the vertical eddy coefficients for momentum and tracers. In addition, minimal background values are prescribed for vertical mixing coefficients to avoid numerical instabilities associated with excessively weak vertical diffusion. Eddy-induced velocity is parameterized following Gent and McWilliams (1990) with a varying coefficient, depending on the growth rate of the baroclinic instability (Treguier et al., 1997). Nevertheless, this parameterisation has no effect on the equatorial dynamics as the coefficient decreases in the tropics and vanishes at the equator.

Lengaigne et al. (2003) used a second order centered advection scheme. One of its weaknesses is that it can produce unphysical oscillations in the profile of an advected quantity with rapidly changing gradients, due to numerical dispersion. To avoid salinity and temperature oscillations in the stratified eastern equatorial thermocline, it is necessary to add diffusion, and the background vertical eddy coefficients are artificially increased
in the top 40 m. In this study, we chose to use a Flux Corrected Transport (FCT) advection scheme (Zalesak, 1979) described and tested in Levy et al. (2001) instead of the usual second order advection scheme. The FCT advection scheme includes some diffusion where sharp gradients are encountered to prevent overshoot. The increase in the background vertical eddy coefficients in the top 40 m is no longer required.

The model is forced from 1993 to 2000 with wind stresses derived by combining TAO data, ERS-1 and ERS-2 scatterometer data, with a smoothed transition to NCEP wind stresses poleward of 60° (Menkes et al., 1998). This stress has been shown to do an excellent job in forcing the model (Vialard et al., 2001; Lengaigne et al., 2003). Atmospheric heat and freshwater forcing fields consist of daily atmospheric temperature NCEP reanalysis (Kalnay et al., 1996) and merged analysis of precipitation (Xie and Arkin, 1996). Heat fluxes are computed using bulk formulae (Goose, 1997). The model starts from an ocean at rest with January potential temperature and salinity fields imposed from the Levitus (1998) climatology. It is spun up for a 3-year period using 1993 forcing fields before starting the interannual simulations.

Four experiments have been performed (Table 1). In the first one (STD), the background vertical eddy viscosity and diffusivity coefficients are artificially increased in the first 40 m, to permit an easier comparison with Lengaigne et al. (2003), and further highlight the benefit of using an FCT advection scheme. For diffusion, they are set to $10^{-4}$ m$^2$ s$^{-1}$ at 5 m depth and decrease linearly to $10^{-5}$ m$^2$ s$^{-1}$ at 40 m depth, while values ten times larger are used for viscosity. It is worth noting that these background values are only reached in regions of low mixing and strong stratification. In regions of strong mixing, the vertical eddy viscosity and diffusivity coefficients are those computed by the TKE equation. Following Lengaigne et al. (2003), the lateral eddy viscosity and diffusivity coefficients are equal to 2000 m$^2$ s$^{-1}$, a value close to the one derived from observations (Bryden and Brady, 1989). This STD simulation is the same as ISOMT in Lengaigne et al. (2003), except that the advection scheme used is FCT instead of Arakawa. The differences induced by these different advection schemes are completely negligible.

The second experiment (RV) is the same as STD, except that the background vertical eddy viscosity and diffusivity coefficients are set to $10^{-5}$ m$^2$ s$^{-1}$ everywhere. Since we use the FCT advection scheme, this reduction does not lead to numerical instabilities in the eastern part of the basin. Such numerical instabilities are indeed observed when using the usual second order advection scheme and $10^{-5}$ m$^2$ s$^{-1}$ for the background vertical eddy viscosity and diffusivity coefficients. Indeed, the FCT advection scheme is slightly diffusive and implicitly adds some diffusion. The background eddy diffusivity coefficients used in the model are thus slightly higher than $10^{-5}$ m$^2$ s$^{-1}$.

The third experiment (RVRI) is the same as RV, except for the lateral eddy viscosity and diffusivity coefficients which are decreased to 1000 m$^2$ s$^{-1}$ in order to test the impact of slightly reducing the mixing along isopycnals. Others experiments with even smaller lateral coefficients (500 and 100 m$^2$ s$^{-1}$) have been performed, but will not be presented in this paper (see discussion in Section 6). The undesirable effect of such a drastic change (one order of magnitude) has been shown in Maes et al. (1997), though with horizontal mixing instead of isopycnal mixing and in a different configuration.

The last experiment (GAL) is the same as RVRI, but with the Galapagos Islands included. A surfacing island is introduced in the bathymetry at 90°W, 0.5°S and the scale factor used for the zonal current at 89°W and 91°W is also divided by 2 at the equator. This operation simulates the presence of islands right on the equator but not north of it.

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<th>Background vertical eddy coefficients (m$^2$ s$^{-1}$)</th>
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3. Impacts on mean currents

3.1. The equatorial undercurrent

The effect of reducing the background vertical turbulent mixing in the upper layers is first evaluated by comparing the zonal currents in the STD and RV experiments. The effect of reducing the lateral turbulent mixing is then evaluated by comparing the RV and RVRI experiments. Fig. 1 shows the longitude/depth diagrams of the 1993–2000 mean equatorial zonal currents for the four experiments, as well as their pairwise differences. Fig. 2 shows the latitude/depth diagrams of the 1993–2000 mean zonal currents at 110°W, which is

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Fig. 1. Left: Equatorial section of mean 1993–2000 zonal currents for the four experiments (from top to bottom: STD (a), RV (b), RVRI (c), GAL (d)). Right: Differences between the different experiments (from top to bottom: RV-STD (e), RVRI-RV (f), GAL-RVRI (g), GAL-STD (h)). Thin lines are for positive values, dashed lines for negative values. Units are cm s⁻¹.
Fig. 2. Left: Meridional section of zonal current from 6°N to 6°S at 110°W for the four experiments (from top to bottom: STD (a), RV (b), RVRI (c), GAL (d)). Right: Differences between the different experiments (from top to bottom: RV-STD (e), RVRI-RV (f), GAL-RVRI (g), GAL-STD (h)). Thin lines are for positive values, dashed lines for negative values. Isopycnals are superimposed in panels (c) and (f). Units are cm s⁻¹.
the region with the strongest impact, for the four experiments, as well as their pairwise differences. In all simulations, the mean equatorial zonal currents are well simulated (Fig. 1), and behave in agreement with observations (Johnson et al., 2002, see their Fig. 2). The EUC starts around 200 m deep in the western part of the basin, strengthens to 156°E, and weakens at 165°E (not shown), before reaching its maximum speed between 155°W and 120°W. It shoals towards the east, following the shoaling of the thermocline, and weakens considerably east of 100°W.

Changing from the STD to the RV run induces effects on the equatorial dynamics that are essentially located in the Eastern Pacific (Fig. 1a, b, and e). The upper part of the EUC is strengthened by 25 cm s⁻¹ east of 120°W. As a result, the upper limit and the core of the EUC are shifted upward. In the same region, the South Equatorial Current (SEC) is trapped and confined to the surface and is decelerated. At 110 °W, the meridional shape of the EUC remains unchanged (Fig. 2a and b), as the whole meridional extension of the EUC is accelerated (Fig. 2e).

The transition from RV to RVRI (Fig. 1f) is associated with a large-scale intensification of the mean EUC in the central and eastern equatorial Pacific. The EUC maximum increases from 96 cm s⁻¹ to more than 100 cm s⁻¹ at 130°W, and the core of the EUC (around 60–150 m depth) is strengthened all along the equator, by up to 13 cm s⁻¹ in its eastern part. Changes in the meridional shape of the mean EUC induced by this reduction in the isopycnal mixing are quite different from those induced by changes in vertical mixing (Fig. 2e and f). Decreasing the lateral viscous and diffusive mixing has only small impacts above 50 m on the equator, but stronger impacts between 50 and 150 m, in the core and in the lower part of the EUC. At these depths, the zonal currents are not only stronger right on the equator, but also slightly weaker off the equator (1/2°S–1/2°N). The shape of the mean EUC is modified, and it becomes narrower meridionally.

In the mixed layer, the mixing coefficients are computed from the turbulent kinetic energy equation model. Decreasing the background vertical turbulent mixing in the first 40 m is only efficient when the background threshold is reached in the model, i.e. when the mixed layer depth is less than 40 m. This is rarely the case in the western and central Pacific, but this is true most of the time in the Eastern Equatorial Pacific Ocean (Wang and McPhaden, 1999; de Boyer Montégut et al., 2004). In this region, changing from the STD to the RV simulation decreases the vertical momentum mixing. It reduces the lost of westward momentum flux by the SEC to the EUC, and the upper part of the EUC is accelerated. However, it is evident in Fig. 1 that the SEC is also decelerated. This deceleration is explained by the westward acceleration due to an increased vertical advection of westward momentum that counteracts the eastward acceleration due to the reduction of the vertical momentum exchange.

As shown by Maes et al. (1997) using horizontal mixing and drastic modifications of the mixing coefficients, and by Lengaigne et al. (2003), the lateral diffusion is a dominant decelerating term in the momentum balance in the core and in the lower part of the EUC. Reducing the lateral eddy diffusivity and viscosity (moving from RV to RVRI) thus naturally induces a progressive acceleration of the EUC all along the equator, which reaches a maximum in the central and eastern Pacific (Fig. 1f). The changes in the meridional shape of the mean EUC can be explained by the relative positions of the isotachs and isopycnals. Lengaigne et al. (2003) showed that when momentum mixing occurs along isopycnals, the lateral diffusion is dominated by its meridional component (90% of the lateral diffusion term). Also, the lateral viscosity flux is proportional to sin(alpha), where alpha is the angle between the isotachs and the lateral mixing direction, the isopycnals. As a consequence, the isopycnal momentum diffusive term has only a small impact on the currents in the top 50 m, where the isopycnals and isotachs are almost parallel from 4°S to 2–3°N (Fig. 2). In contrast, the isopycnals encounter strong zonal currents gradients between 50 and 150 m. Reducing the meridional mixing of momentum at these depths has thus a stronger impact on the equatorial currents.

The effect of introducing the Galapagos Islands in the bathymetry of the model are finally evaluated by comparing the GAL and RVRI experiments. It is known that the Galapagos Islands provide a topographic barrier for the equatorial currents. As the EUC encounters the Galapagos Archipelago, it is split into two branches that continue toward South America (Lukas, 1986; Kessler, 2006). It has also been shown that the topography of the Galapagos Islands induces a local upwelling on the western side of the islands (Eden and Timmermann, 2004; Brentnall, 1999) and modifies the intensity of the SEC. In our simulations, it appears from Fig. 1d and g that the impact of Galapagos Islands on the zonal currents is strictly restricted to the east and to a few degrees west of the islands (located at 0.5°S, and 90°W) as found by Eden and Timmermann
Because of the new topography, the strength of the EUC and the SEC are strongly reduced locally by more than 15 cm s\(^{-1}\). Upstream of the Galapagos Islands, the impact is small.

Finally, reducing both the vertical and lateral turbulent mixing induces a substantial change in the simulated EUC in the central an eastern Pacific (Fig. 1h), and the mean speed of the EUC core increases by more than 29 cm s\(^{-1}\), whereas introducing the Galapagos Islands seems to have only a limited impact upstream at the low resolution we are using. Whether these changes are really improving the realism of the model will be examined in Section 3.3.

3.2. Associated mean 3-D circulation

3.2.1. Description of the mean simulated tropical cell (TC) branches

In our simulations, decreasing the diffusion coefficients greatly strengthens the modelled EUC. It is suspected that it also induces large changes in the 3-D circulation in the eastern Pacific, and in the TC strength. How sensitive are the TC to changes in the vertical and horizontal diffusion parameterisations? How do they interact with the Galapagos Islands? These questions are now explored.

The structure of the TC 3-D circulation is first described with the STD experiment. The TC are usually referred to as the mean 2-D circulation which is depicted in a meridional cross-section near the equator. We aim to enlarge this zonal mean vision, especially in the eastern equatorial Pacific, where the winds and the TC circulation are specifically asymmetric. Zonal winds are weak and the northward meridional component of the winds becomes important. As the upwelling, which becomes shallower approaching the coast (with a depth of maximum upwelling evolving from \(~60\) m at 120°W to \(~20\) m at 90°W, not shown), the TC and the EUC core drift slightly south of the equator (Figs. 2a, 4a, 3c). Associated with this drift, an anomalous meridional cell has been identified in numerical studies (Kessler et al., 1995). It occurs at the equator (northward at the surface, southward in the pycnocline), and modifies the symmetric cells caused by the zonal wind. However, in the STD experiment, this asymmetric cell is curiously not observed west of 90°W, and the equatorial meridional current flows slightly southward against the northward wind. It is possible that the 10 m vertical resolution may not be sufficient to simulate this very shallow cell.

Asymmetries in the STD simulated transports are also evident in Figs. 3–5. Mean surface poleward currents are observed west of \(~90°\)W at 2°N, and west of \(~85°\)W at 2°S (Fig. 3), but their extension is seasonally dependent. The transport of the TC surface branch, located at a deeper surface layer, is nearly twice as large at 2°S than at 2°N. The meridional pycnocline convergence is much smaller in the southern hemisphere. In the northern hemisphere, the presence of the ITCZ induces a stronger diapycnal downwelling at 2–5°N of about 1–1.5 m/day, increasing the convergence. Its zonal structure is also different: it extends almost to the eastern boundary at 2°N, whereas it is located only west of \(~100°\)W or west of 90°W at 2°S (Fig. 3), depending on the season. The preferential southward outflow of EUC water in the southern surface layer, and southward inflow into the EUC in the northern pycnocline are well explained by the hemispheric asymmetries in the zonal wind stress, and in the wind stress curl (Blanke and Raynaud, 1997).

East of 100°W–90°W, a poleward subsurface divergence is observed, resulting from a deviation of the EUC waters. The TC structures are thus complex as we approach the coast of South America. It is difficult to talk of a TC east of the Galapagos Islands location, where the mean surface and thermocline currents reverse, so we will define and study the TC as west of 90°W in the following sections.

3.2.2. Sensitivity of the 3-D circulation to vertical and lateral diffusion and to the presence of the Galapagos Islands

Fig. 4 shows the latitude/depth diagrams of the 1993–2000 mean meridional currents averaged from 110°W to 90°W for the four experiments, as well as their pairwise differences. Fig. 5 gives transports estimates in the different experiments. When we reduced the background vertical turbulent mixing in the upper layers (Section 3.1), we showed that the EUC in the eastern Pacific became narrower and much stronger in RV than in STD. We also observe in RV an increase in the equatorial upwelling, in the diapycnal downwelling at 4°N, in the surface meridional divergence, in the pycnocline convergence and in the SEC transport (Figs. 4e and 5). The currents in all TC branches are thus significantly increased. For example, the vertical current at the equator is increased by 40% from STD to RV, and the surface meridional currents are increased by more than 50%
east of 110°W, and by 100% east of 100°W. However, the spatial structure of these currents is also changed: meridional poleward divergence is on a shallower surface layer in RV compared to STD (Fig. 4a, b and e), especially in the southern hemisphere where it extends quite deeply. This is because changes in the vertical mixing affect the Ekman depth, and thus the depth at which the wind stress penetrates into the ocean. The spatial structure of the upwelling is also changed, related by mass continuity to the variations in divergence. The upwelling extension is more concentrated on the equator, with a slightly tighter meridional width. What happens is that the increase in meridional velocity is stronger close to the equator in RV compared to STD. Therefore, there is an anomalous convergence of surface meridional velocity poleward of 2°, and the meridional width of the upwelling is reduced. Hence, if the TC currents are greatly affected by the changes in the vertical mixing, the transports are less affected: the surface poleward transport is increased by 27% at 2°N, and by only 5% at 2°S (Fig. 5), whereas vertical transport between 110°W and 90°W at 40 m depth increases from 10.4 Sv in STD to 12.3 Sv in RV (18%).

The transition from RV to RVRI is associated with much smaller TC changes. Pycnocline convergence, equatorial upwelling and surface divergence slightly decrease (Figs. 4f and 5). Unlike the EUC transport, the TC are not very sensitive to small changes in the horizontal parameterisation.

Finally, in the GAL experiment, only small changes occur in the mass transports for all of the branches of the TC (Figs. 4g and 5). Qualitatively, the Galapagos Islands make the EUC split into two branches. Their presence diminishes the EUC and the upwelling east of the islands, but increases the upwelling to the west. The changes induced in the surface meridional currents are negligible, except for local flow turning around

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Fig. 3. Mean 1993–2000 meridional current at 2°N (a) and 2°S (b), for the STD experiment. Units are cm s⁻¹. Some isopycnals (in kg m⁻³) are superimposed. Mean 1993–2000 vertical current at 40 m (c). Units are m day⁻¹. Thin lines are for positive values, dashed lines for negative values.
Fig. 4. Left: Meridional section of meridional current from 6°N to 6°S averaged between 110°W and 90°W for the four experiments (from top to bottom: STD (a), RV (b), RVRI (c), GAL (d)). The arrows represent the meridional and vertical transports (in Sverdrups). Their directions have been adjusted to the aspect ratio of the figures. Right: Differences between the different experiments (from top to bottom: RV-STD (e), RVRI-RV (f), GAL-RVRI (g), GAL-STD (h), contours every 0.5 cm s\(^{-1}\)). Thin lines are for positive values, dashed lines for negative values. Units are cm s\(^{-1}\).
the Galapagos Islands. Hence, it appears that the presence of the Galapagos Islands has no effect on the TC transport. Alternatively, it is interesting to note that the subsurface divergence is significantly increased at the depth of the EUC core to the west of the islands (Figs. 4g and 5).

In addition to the TC transports, we wish to determine the sensitivity of the EUC termination to the different parameterisations. First of all, it is shown that the EUC transport reaching the South American coast is greater when there is decreased vertical and lateral mixing (Fig. 5). Although the EUC deceleration by entrainment (Pedlosky, 1988) is important, it is not sufficient to explain the EUC termination, and an inertial collision at the coast is also needed. To understand which pathways the EUC waters are taking, a poleward subsurface divergence has been computed east of 100ºW as indicated in Fig. 5. For this subsurface divergence, about half the water is flowing to the North, and half to the South. For the southward flowing waters, we differentiate between two branches. The first corresponds to the branch of the EUC that flows along the equator as far as South America, and then turns southward along the coast. The second is associated with the branch that flows southeastward to reach the coast at around 5ºS (Lukas, 1986). When the vertical mixing is decreased, the transports of both the northern and southern branches are increased equally. When the lateral mixing is reduced, the transports of the southern branch are increased more than for the northern branch, due to an intensification of the coastal undercurrent. Interestingly, the presence of the Galapagos Islands appears to modify the splitting of the EUC in the southern hemisphere: the coastal undercurrent greatly decreases when introducing the Galapagos Islands bathymetry, which in turn increases the southward flow away from the coast. Therefore, the presence of the Galapagos Islands in the simulation leads to a change in the pathways of the EUC waters, which diverge southward and join the coastal currents. The EUC branch that reaches the coast on the equator and then turns southward is dramatically reduced, whereas the branch that diverges southward before reaching the coast is significantly increased.
3.3. Comparison with observations

To evaluate the realism of our simulations, the simulated currents of the three experiments are now compared to observations. In the Equatorial Pacific, zonal ocean currents are available from the TAO array. The TAO array (Hayes et al., 1991; McPhaden et al., 1988) consists of approximately 70 moorings in the Pacific Ocean measuring surface and subsurface temperature, surface winds and others meteorological parameters. At some equatorial sites (147°E, 156°E, 165°E, 170°W, 140°W and 110°W), ADCP (Acoustic Doppler Current Profilers) and/or mechanical current meter are also deployed and measure ocean currents down to 300 m. Owing to instrumental failure, the time series have missing values at some times and depths. Several steps are thus needed to make an accurate comparison between data and simulations. First, gaps are filled whenever possible using regression relations based on data from adjacent depths, with a procedure similar to that described by Johnson and McPhaden (1993). Then, vertical spline interpolations are performed every 5 m for the zonal currents (Izumo, 2005). Finally, the daily current data are averaged into 5-day bins, and the simulated zonal currents are extracted at the TAO mooring locations at every time step where data are available during the 1993–2000 period.

Fig. 6 shows the resulting mean vertical profiles of the TAO and simulated zonal currents at 140°W and 110°W showing the EUC structure down to 220 m, for the STD, RV and RVRI experiments. Since the effect of adding the Galapagos is negligible here, the GAL profile is not included. The comparison of the STD simulation with all TAO moorings has already been presented in Lengaigne et al. (2003), and shows a very good agreement between the model and the data. As this paper focuses on the eastern Pacific, the vertical profiles are only shown at the two easternmost moorings, at 140°W and 110°W. First of all, it appears that the mean vertical current structure is well captured by the four simulations. It is also evident that important improvements are induced by the different choices of turbulent mixing coefficients. In the STD case, the strength of the EUC is slightly underestimated in the model at 140°W and 110°W, and the core of the simulated EUC is weaker than in reality. Reducing the background vertical mixing in the first 40 m (moving from STD to RV) improves the realism of the simulated EUC at 110°W in its upper part, even if the simulated SEC still remains too deep. The reduction in the lateral mixing coefficients (moving from RV to RVRI) also greatly improves the comparison with data, especially at 110°W. At this mooring, the core and the lower part of the simulated EUC are accelerated by 19 cm s⁻¹ and are much closer to the observations. At 140°W, they are accelerated by 10 cm s⁻¹.

We compared the zonal simulated currents on the equator. However, as the vertical and meridional profiles of zonal currents are also modified in our experiments (Fig. 2), it is worth computing the zonal mass transport.

![Fig. 6. Mean 1993–2000 zonal current profiles of TAO data (thick line) and of experiments STD (thin line), RV (dotted line), and RVRI (dashed line). Comparisons at two moorings (140°W and 110°W). Units are cm s⁻¹.](image-url)
Table 2
Mean 1993–2000 EUC transport and kinetic energy right at the equator at 110°W, from the TAO mooring and the different simulations

<table>
<thead>
<tr>
<th></th>
<th>$U_{EUC/eq}$ (m$^2$ s$^{-1}$)</th>
<th>$KE_{EUC/eq}$ (J/kg/m)</th>
<th>$U_{EUC}$ (Sv) (2.5°S–2.5°N)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>110°W, 0°</td>
<td>110°W, 0°</td>
<td>95°W</td>
</tr>
<tr>
<td>TAO or ADCP</td>
<td>101</td>
<td>40.7</td>
<td>21.9</td>
</tr>
<tr>
<td>STD</td>
<td>73.2</td>
<td>23.1</td>
<td>19.2</td>
</tr>
<tr>
<td>RV</td>
<td>83.0</td>
<td>30.1</td>
<td>21.4</td>
</tr>
<tr>
<td>RVRI</td>
<td>94.7</td>
<td>37.8</td>
<td>22.2</td>
</tr>
<tr>
<td>GAL</td>
<td>95.2</td>
<td>37.7</td>
<td>22.3</td>
</tr>
</tbody>
</table>

The temporal means of EUC transport at 110°W and 95°W estimated over the whole meridional extent of the EUC, are also presented (two right columns), from CTD/ADCP data (from Johnson et al., 2002) and the different simulations.

The simulated EUC mass transports computed on the equator, and between 2.5°S and 2.5°N, are given in Table 2, together with the TAO EUC mass transports on the equator and with the EUC mass transport in the 2.5°S–2.5°N band deduced from the ADCP data (Johnson, 2001). The EUC kinetic energy is also given. The method used is the same as in Izumo (2005), and is detailed in the Appendix. When the vertical and lateral mixing are decreased, the EUC becomes stronger and narrower at 110°W. In the 2.5°S–2.5°N latitudinal band, the EUC mass transport increases by 2.2 Sv (10%) from STD to RV and by 3.0 Sv from STD to RVRI (16%). At the equator, there is a stronger increase of 12% from STD to RV and ~25% from STD to RVRI, and an important increase in kinetic energy of 30% from STD to RV and ~60% from STD to RVRI (Table 2). This contributes to the zonal extension of the EUC in the extreme east of the basin, through increased kinetic inertia.

With the changes induced by our sensitivity experiments, the profiles of the mean zonal currents are very well captured. The position and the strength of the EUC core are very realistic. However, some deficiencies still remain even in the most realistic RVRI simulation. At 110°W in the surface layers, the mean simulated SEC is too strong and too deep compared to the observations. It is important to identify which season has the greatest differences, in order to understand their physical source.

4. Seasonal variability

4.1. The EUC seasonal cycle

In the eastern Pacific, the EUC transport is strongly varying at seasonal timescales, resulting from both wind variations and wave propagations. Fig. 7 shows the mean seasonal cycle of the observed and modelled STD and RVRI zonal currents at 110°W, at 35 m depth (the TAO time series at 15 and 25 m depth are more gappy). At this depth and longitude, RV and GAL zonal currents are almost identical to RVRI zonal currents and are not included for the clarity of the figure. At 110°W, mean surface currents are small. This feature has also been noted by Kessler (2006) and Johnson et al. (2002), analyzing both drifters and ADCP data. It can be explained by the annual surfacing of the EUC during boreal spring. At this time of year, the trade winds weaken, the EUC strengthens and surfaces so that the surface currents can reach 30 cm s$^{-1}$, inducing a near zero surface mean current (Yu and McPhaden, 1999; Yu et al., 1997). Fig. 7 shows that this seasonal feature is not well simulated in the STD experiment, but is quite well captured in the RV and the RVRI experiments. In the STD experiment, the vertical background mixing is indeed stronger than in RVRI. During boreal spring, the mean mixed layer depth is very shallow (Wang and McPhaden, 1999) and the background values are reached: an increase in the vertical background mixing has an important impact on the simulation during this period. Noticeable differences occur between TAO data and modelled 35 m currents during boreal autumn (Fig. 7). This is even more obvious when examining the surface currents (not shown). During these periods, the trade winds and the westward currents are strong, and significant tropical instability waves are generated (e.g. Ryan et al., 2002). The Richardson number is often subcritical in the surface layers, which means that the background values for the mixing coefficients are rarely reached, and moving from STD to RV during these cold periods leads to a smaller improvement. For the same reason, further reduction of the background vertical mixing coefficients is not of great help.
4.2. The TC seasonal cycle and the link with the EUC seasonal cycle

To our knowledge, the seasonal cycles of all of the branches of the TC in the eastern Pacific and their link with the EUC transport have never been precisely explored, and we propose to investigate how they participate in the EUC termination. Fig. 8 shows the seasonal cycles of the TC branches, as well as the seasonal cycles of the subsurface divergent transports in a 2°S–2°N/110°W–80°W box, the southward coastal undercurrent at 2°S, and the EUC and SEC transports at 110°W. The seasonal cycle of the local wind is also plotted. The surface divergence triples from its minimum in March (6 Sv in RV) to its maximum in July (17 Sv), in phase with zonal wind stress variations. This is coherent with the conclusions of Poulain (1993) and Meinen et al. (2001) obtained with data from drifters. The upwelling is well correlated with a one month lag, increasing from its minimum in April (5.5 Sv in RV) to its maximum in August (20.5 Sv). The pycnocline convergence has a similar evolution (from 1 Sv in March–April to 9–10 Sv in August–November).

The seasonal cycles of the TC meridional and vertical transports are thus not in phase with the EUC transport, is seasonal cycle. In the eastern Pacific, the TC transports are well correlated with the local wind stress, and respond directly to its seasonal variations. On the contrary, they seem to be dynamically independent of the EUC transport’s seasonal cycle, which depends both on local forcing and remotely forced equatorial waves (Yu and McPhaden, 1999). Interestingly, it appears that the seasonal cycle of the subsurface poleward divergence (Fig. 8c) is in phase with the EUC transport seasonal cycle. This is not the case in the central Pacific (Blanke and Raynaud, 1997), where the EUC mass excess during its seasonal maximum is supplied by meridional convergence and expelled by divergence. The TC are not locally feeding the EUC mass excess. On the contrary, the EUC mass surplus is evacuated poleward in the subsurface layers, and is notably feeding the PCUC (Peru Chile UnderCurrent).

This dynamical independence is confirmed by the changes induced by our sensitivity experiments. When the lateral diffusion is diminished, the mean EUC transport increases whereas the mean TC transports decrease. Clearly the seasonal EUC mass excess is not supplied by the TC and does not feed them. The situation is different when the vertical mixing is diminished. When moving from the STD to the RV case, increases in both the EUC and the TC transports are observed. To understand if these increases are related, we examine their seasonal timing in both simulations. The EUC transport increase from STD to RV is maximum in April–May–June, when the EUC surfaces and when its transport is the highest. It is negligible from August to October, when the EUC is deeper and its transport is smaller. On the contrary, the increases in mass transport of the pycnocline convergence, upwelling and surface divergence from STD to RV are greater in August–October than in April–May–June. What happens is that during boreal spring, moving from STD to RV leads to an important increase in the surface eastward current, and the consequent decrease in the geostrophic divergence counteracts the increase in Ekman divergence. During other seasons, the EUC is deeper, and less affected by the reduction of vertical mixing in the surface layers. The surface zonal current remains unchanged, and the Ekman divergence is increased because of the reduction of vertical mixing. Therefore, it appears that the
Fig. 8. Seasonal cycles of EUC and SEC transports at 110°W, surface divergence, pycnocline convergence, upwelling at 30 metre and subsurface divergence transport from 110°W to 80°W and 2°S to 2°N for STD (thin line) and RV experiments (dashed line). Units are in Sverdrups. See Appendix for the calculation details. The amplitude of the zonal stress averaged from 110°W to 80°W and 2°S to 2°N is also plotted (in N m⁻²).
increase in transport in the EUC and TC from one simulation to the other are largely independent, as are the seasonal variations of EUC and TC transports. Conversely, the increase of subsurface divergence from STD to RV is maximum in April–May and minimum in August–September, which is in phase with the EUC transport increase. The EUC transport increase induced by the reduction of vertical mixing is expelled by the subsurface poleward divergence.

To summarize, our study showed that the 3-D circulation is enhanced in the eastern Pacific when vertical diffusion is reduced, with an increase in the EUC, in the surface meridional divergence, in upwelling, in pycnocline convergence and in subsurface divergence near the coast of South America. However, these increases in TC meridional and vertical transports appear to be independent of the EUC transport increase, as shown by the distinct timing of the seasonal changes in the different simulations. Therefore, the TC increases are not directly related to the EUC transport increases. Conversely, the EUC transport increase appears to feed the subsurface divergence increase when approaching the South America coast. When the isopycnal diffusion is reduced, the TC changes are opposite and smaller, with a slight slowing down of TC transport. The EUC transport increase thus feeds the subsurface divergence. Finally, the TC transports remain largely unchanged when introducing the Galapagos in the model bathymetry. However, the splitting of the EUC appears to be more important at depth.

5. Impacts on surface temperature

In the four sensitivity experiments, the vertical and lateral mixing coefficients for tracers were also changed along with those for momentum. Therefore, we expect changes in the mean simulated temperature and salinity fields. Moreover, modifying the strength of the EUC will certainly affect the SST in the eastern Pacific and the vertical temperature profiles in the equatorial region.

Fig. 9 shows longitude–latitude maps of 1993–2000 mean equatorial temperatures for the STD experiment, the differences between the four experiments and the Reynolds SST for comparison (Reynolds and Smith, 1994). It also presents an equatorial zoom of the Reynolds and modelled SST. It is evident that the mean SSTs are realistically simulated in the four experiments. The Warm Pool and the Eastern Pacific are too warm by about 1 °C, but the gradient between the Warm Pool and the cold tongue are very well reproduced.

Differences in sea surface temperatures between the STD and RV experiments (where the background vertical mixing coefficient has been reduced) are clearly seen in the eastern Pacific (Fig. 9d). Right on the equator, the reduction in vertical mixing (RV) shows mean SSTs which are cooler by more 0.5 °C east of 110°W, and up to 0.65 °C in the far eastern Pacific. This cooling takes place during the whole year, but is more pronounced during the boreal spring (when a 1.4 °C mean seasonal cooling is observed, not shown) when the EUC surfaces. This improves the realism of our simulation (Fig. 9a–c). This cooling is important since it has been shown that a variation of ±0.5 °C at 110°W–0° can cause significant variations in sensible and latent heat flux (Cronin and Kessler, 2002). These mean SST changes can modify the coupled ocean–atmosphere system, by their impact on the winds and cloud structures.

The explanation for this cooling is not simple. It appears to be more related to dynamics than to a reduced mixing of tracers. During boreal spring, the annual surfacing of the EUC is better simulated, and the temperatures are cooler. Moreover, the EUC mass transport is increased in the RV experiment and there is more water arriving at the South American coast. The vertical and meridional circulation associated with the EUC also contributes to this cooling. East of 120°W, the Tropical Cells are reinforced, with increased surface divergence, pycnocline convergence and upwelling in the first few metres. This results in stronger mixing and cooler SSTs. As noted for the zonal currents, the advection processes counteract and are much more efficient than the diffusion processes.

Moving from RV to RVRI (reducing the lateral mixing) induces by contrast an equatorial warming of about 0.2–0.25 °C (Fig. 9e). Knowing that the model is already too warm in the eastern Pacific, this is not an improvement. The upwelling and the poleward divergence are decreased in the surface layers (Fig. 4) and this partly explains the slight warming (~0.2–0.25 °C) of the SST observed in RVRI compared to RV. A rough estimate with a simple heat transport analysis indicates a ~0.1–0.15 °C warming due to this decrease. Moreover, it has been shown by Pezzi and Richards (2003) that decreasing the lateral mixing in the OPA OGCM increases the activity of Tropical Instability Waves (TIWs), resulting in a warming of the cold tongue.
The eddy kinetic energy has been computed in the central and eastern equatorial Pacific, and it appears that TIWs are indeed stronger in the RVRI than in the RV experiment.

Finally, the GAL experiment (introducing the Galapagos Islands in the bathymetry) shows a strong impact on the SST around the islands. A local upwelling west of the Islands is observed, and the SST is cooler by more than 1.4 °C on the equator and by about 0.2 °C at 4°N (Fig. 9f). East of the islands, it is warmer by 1 °C. This local warming seems too strong when compared to the local warming in Reynolds SST. However, the local cooling seems realistic, and may impact on local air–sea interaction and the local ecosystem (Eden and Timmermann, 2004). Including the Galapagos Islands may be important for future simulations.

Similar changes are seen in the salinity fields, with the opposite sign, since the EUC brings colder and saltier water to the eastern Pacific. These salinity changes contribute as much as the temperature changes to modifying the density field in the eastern Pacific. The induced stratification modifications may be important for the dynamics and coupling in the eastern Pacific.

6. Conclusions and discussion

In this paper, sensitivity experiments have been performed in ORCA2, the global 2° configuration of the OPA OGCM, to better understand the 3-D circulation in the eastern equatorial Pacific. They allowed us to determine the structure of the Tropical Cells in the East, defined as west of the Galapagos Islands, and appear to be useful tools to investigate the link between EUC mass transport and its associated meridional and
vertical circulation which together govern its termination. They also allow us to improve the realism of the simulated equatorial currents in the region. The coefficients involved in the parameterisations of the sub-grid-scale physics have been modified, and Galapagos Archipelago has been added in the bathymetry. An FCT advection scheme has been used in our simulations instead of a second order centered advection scheme.

The FCT advection scheme appears to be a very good option to use as it allows us to set smaller and more realistic mixing coefficients, without inducing numerical instabilities. This is of great importance in regions with a strongly stratified ocean, such as the eastern equatorial Pacific. Indeed, it has been shown in this paper that reducing the vertical eddy coefficients leads to a great improvement in the EUC simulation. The depth over which the wind stress penetrates the water column has also been modified, and this also induces changes in the meridional and vertical circulation. The 3-D circulation is thus enhanced in the eastern Pacific, with an increase in the EUC, the surface meridional divergence, the upwelling, the pycnocline convergence and in the subsurface divergence near the coast of South America. Physically, reducing the vertical background mixing allows the mean vertical advection of momentum and temperature to increase and work efficiently when necessary. During boreal spring, the surface of the EUC is notably better simulated.

Reducing the lateral mixing coefficient to 1000 m$^2$ s$^{-1}$ instead of 2000 m$^2$ s$^{-1}$ (both values being in a range of physical values) allows the EUC to be more inertial in its core and in its lower part. The mean speed of the EUC increases by more than 29 cm s$^{-1}$ in its core at 110$^0$W, and becomes therefore very close to the observed TAO zonal currents at this mooring. The TC transports are conversely not noticeably modified, and the EUC mass surplus is expelled poleward in subsurface.

Adding the Galapagos Archipelago does not induce significant changes in the upstream equatorial circulation, and thus produces no changes in zonal currents and TC strength at 110$^0$W. However, the presence of the islands leads to a substantial change in the pathways of EUC waters that join the coastal currents, which may be important for water masses properties east of the islands and along the coasts of South America. The changes introduced also affect the SST and the SSS in the eastern Pacific. They could be of importance for simulations coupling the oceanic GCM with an atmospheric GCM.

Interestingly, the EUC and TC branches transports are not in phase and appear to be dynamically independent at seasonal timescales. A lagrangian analysis has shown that the TC explain a large part of the seasonal variability of the EUC transport in the central Pacific (Blanke and Raynaud, 1997), whereas our study demonstrates that TC are not locally feeding the seasonal EUC mass excess in the eastern Pacific. On the contrary, the spring EUC mass surplus is evacuated poleward in the subsurface layers, and is notably feeding the Peru-Chile Undercurrent.

Some deficiencies in the simulated equatorial currents still remain in the eastern Pacific. The simulated SEC is too strong and too deep at 110$^0$W, 0$^0$, whereas the TAO observed currents indicate a weak mean surface current. Additional experiments have been undertaken to try to resolve this discrepancy, without success. The background vertical mixing coefficients were decreased even more. Lateral mixing coefficients were reduced to 500 and 100 m$^2$ s$^{-1}$. Although choosing a value of 500 m$^2$ s$^{-1}$ seems physically justifiable, using a value of 100 m$^2$ s$^{-1}$ induces an additional deepening and an acceleration of the EUC that is unrealistic compared to the observations. For example, at 110$^0$W, the EUC is 20 m deeper and 18 cm s$^{-1}$ stronger than observed. Moreover, in all cases, the surface SEC is still too strong at 110$^0$W compared to the TAO observations. Similar conclusions have been obtained by Pezzi and Richards (2003). We are thus reaching the limit of our sensitivity exercises.

Two physical hypotheses can be proposed to explain this remaining deficiency. The surface currents are not well simulated during la Nina and during summer and autumn periods, when mixing, instability waves, easterly winds and surface currents are stronger (Yu et al., 1997). Thus, it is possible that the lateral mixing is not adequately parameterized, and that we need an improved parameterisation which is intermittent in time and space (Pezzi and Richards, 2003; Richards and Edwards, 2003). This would require further observations, and diagnostics of higher resolution models. On the other hand, as shown by Yu et al. (1997), the annual cycle of the upper circulation in the eastern equatorial Pacific is closely related to the local zonal and meridional winds. Therefore, the discrepancy between the modelled and the observed zonal currents in the Eastern Pacific is not related to the ocean model physics but rather to the forcing. It is worth noting that the excellent quality of our simulations should be in part attributed to the quality of the wind product used to force the model. NCEP or ERS wind stresses are too weak compared to the TAO wind stresses and lead to a less energetic and less realist
EUC. The TAO-ERS wind stresses used in this study appear to do a better job in forcing the model (Vialard et al., 2001; Radenac et al., 2001). Furthermore, the OCM should ideally be forced by wind stresses that depend on the oceanic surface currents (Pacanowski, 1987; Luo et al., 2005). Such a computation will be investigated in a forthcoming study.

With the improvements brought by this paper, we now have the optimal and most realistic simulations of the equatorial currents achievable in the ORCA2 resolution of the OPA model to date. Our simulations do an excellent job in simulating realistic equatorial currents. To reach this objective, we have changed the advection scheme and the physical parameterisations, always maintaining these within reasonable and physical limits. Further improvements seem hard to achieve with the $2^\circ$ resolution model, and will require the implementation of new parameterisations. They will rely on lateral physics varying with space and time, taking into account large lateral mixing in the Western Pacific (Richards and Banks, 2002) as well as better parameterisation of the enhanced mixing associated with the tropical instability waves.

Acknowledgements

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Appendix. Criteria used to compute the EUC transport and the TC branches transports

(a) Definitions of the various branches of the TC:
The TC meridional transports are computed at 1.8°N and 1.8°S, where they are at their maxima. To define the TC for interannually varying currents, temperature and salinity, these fields are temporally and spatially filtered (using 3 months and 6° longitude Hanning filters to remove periods lower than 1.5 months and wavelengths lower than 3°. This in particular removes the TIW activity). The surface meridional poleward divergent transport is then computed by integrating $V_{\text{filtered}}$ over the longitude range selected and between the surface and 100 m depth where the following criteria are valid:

$$\text{At 1.8°N: } V_{\text{filtered}} > 0 \text{ and } \sigma_{\text{filtered}} < 23 \text{ kg m}^{-3}$$

$$\text{At 1.8°S: } V_{\text{filtered}} < 0 \text{ and } \sigma_{\text{filtered}} < 24.75 \text{ kg m}^{-3}$$

where $\sigma$ is the potential density. Note that different $\sigma$ ranges are chosen for 1.8°N and 1.8°S because of the hemispheric asymmetry in isopycnal depth (Fig. 3).

In the same way, subsurface meridional poleward divergent transport is computed by integrating $V_{\text{filtered}}$ over the longitude range selected where the following criteria are valid:

$$\text{At 1.8°N: } V_{\text{filtered}} > 0 \text{ and } 23 \text{ kg m}^{-3} < \sigma_{\text{filtered}} < 26 \text{ kg m}^{-3}$$

$$\text{At 1.8°S: } V_{\text{filtered}} < 0 \text{ and } 24.75 \text{ kg m}^{-3} < \sigma_{\text{filtered}} < 26 \text{ kg m}^{-3}$$

The pycnocline meridional equatorward convergent transport is similarly computed by integrating $V_{\text{filtered}}$ over the longitude range selected and between the surface and 200 m depth where the following criteria are valid:

$$\text{At 1.8°N: } V_{\text{filtered}} < 0 \text{ and } \sigma_{\text{filtered}} < 26 \text{ kg m}^{-3}$$

$$\text{At 1.8°S: } V_{\text{filtered}} > 0 \text{ and } \sigma_{\text{filtered}} < 26 \text{ kg m}^{-3}$$
(b) EUC definition:

For the comparison with the TAO data showed in Table 2, we use the criteria chosen by Izumo (2005), to accurately define the EUC even when the zonal current and temperature fields are varying interannually. The zonal current right at the equator is integrated over the 25–250 m depth range where the following criteria is valid:

\[ U > 0 \text{ and } T < 27 ^\circ \text{C} \text{ and } T - T(15 \text{ m}) < -0.1 ^\circ \text{C}. \]

The comparison with ADCP/CTD sections from Johnson et al. (2002) can be made only with mean currents. Thus, the simple criteria \( U > 0 \) is sufficient when integrating over the 0–400 m depth range and over the adequate latitude range (3\( ^\circ \)N and 3\( ^\circ \)S at 125\( ^\circ \)W, between 2.5\( ^\circ \)S and 2.5\( ^\circ \)N at 110\( ^\circ \)W and between 2.5\( ^\circ \)N and 6\( ^\circ \)S at 90\( ^\circ \)W).

References


Lukas, R., 1986. The termination of the equatorial undercurrent in the eastern Pacific. Prog. Oceanogr. 16, 63–90.


