

Analysis of mean solution characteristics of an eddy-resolving numerical model simulating tropical instability waves in the Pacific Ocean

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Abstract. Using an eddy-resolving ocean general circulation model called INMIO, we study mean thermohydrodynamic conditions in which the equatorial convergence of eddy meridional heat transport (MHT) is formed by tropical instability waves (TIWs). In nature the TIWs warm the well-known cold tongue of the East Pacific waters. The model has sufficient resolution to reproduce the TIWs, but in the vicinity of 110°W it shows significantly underestimated convergence values compared to estimates of direct observations. Analysis of mean temperature distributions in comparison with the WOA09 climatology suggests that the underestimation of the eddy MHT convergence is a model bias, rather than an error of the observational data.

1. Introduction

Eddies and other mesoscale processes play a fundamental role in the formation of the World Ocean climate, determining the local conditions in which the background large-scale circulation develops. Satellite altimetry data show that eddies are widespread throughout the ocean, but their contribution to the transport processes is significant only in certain regions, mainly associated with strong currents and their interactions. These include, for example, the tropics, where tropical instability waves develop on the shears between currents and countercurrents. Under these conditions, the eddy transport may show interesting features, such as horizontal heat transport in the direction of the temperature gradient (i.e. from cold waters to warm ones). Often it is opposed to the transport of the mean flow and has comparable magnitudes [1].

In recent years, due to the development of computational systems, physical methods and numerical schemes, the general circulation models became able to realistically reproduce the large-scale state and variability of the ocean, provided that the boundary conditions are correctly set. A great deal of work has been devoted to the study of the role of mesoscale eddies, in which emphasis is placed on various effects on the scales of the World Ocean and its individual basins, for example, the vertical heat transfer ([2] and references therein), the formation of currents' structure [3], meridional transport of heat and fresh water [4], restratification and deep convection [5], etc. Some studies directly relate the reproduction of eddy dynamics to important qualitative characteristics of models, such as the model climate drift [6] and the separation points of western boundary currents [7].



In [8], the distributions of eddy meridional heat transport (MHT) in the World Ocean and its basins were analyzed by performing a numerical experiment with the eddy-resolving ocean general circulation model INMIO. Eddy characteristics of several circulation elements, in particular, of the main western boundary currents and tropical instability waves (TIWs), were investigated. The goal of this paper is to study the conditions under which the model eddy MHT of the Pacific TIWs is formed. We continue to analyze the mean equatorial Pacific Ocean thermohydrodynamic characteristics of the numerical solution, which was first presented in [8].

2. Model, numerical experiment setup, and eddy MHT calculation

2.1. INMIO World Ocean model

Since the publication in 2000 of the study [9], in which the quality of reproduction of the sea surface height mean-square variability was analyzed, it has become almost a common standard for the model horizontal resolution to be not lower than 0.1° in order for the model to be considered eddy-resolving. Therefore, this work is a step towards the creation of an effective ocean general circulation model suitable for carrying out computations over a wide range of spatial and temporal scales: from long-term climate studies to operational forecasts with high resolution.

The current version of the INMIO model has been developed at the Marchuk Institute of Numerical Mathematics RAS and the Shirshov Institute of Oceanology RAS. It solves the 3DPEM system of ocean dynamics equations with the Boussinesq and hydrostatics approximations by means of the finite-volume method on a B-type grid in vertical z -coordinates with the nonlinear kinematic condition on the free ocean surface. The standard 0.1 -degree grid configuration WOM-0.1 is built on a three-polar horizontal grid with the poles located in Siberia, Western Canada, and on the geographical South Pole. The vertical discretization consists of 49 horizons with steps varying from 6 m in the upper layer of the ocean to 250 m in deep waters.

Except for vertical turbulent mixing, explicit time schemes are applied for all non-stationary processes. This made it possible to effectively implement the model on high-performance distributed-memory computers under control of the Compact Modeling Framework software [10]. Some more detailed characteristics of the model are given in [11,8].

2.2. Numerical experiment setup

Calculations of the eddy-resolving experiment were carried out with a tracer diffusion coefficient equal to $100 \text{ m}^2/\text{s}$ at the equator and scaled towards the poles proportionally to the square root of the grid cell area. In the momentum equations, the artificial Laplacian viscosity was not used, and to ensure numerical stability we implemented a biharmonic filter with a coefficient of $-18 \cdot 10^9 \text{ m}^4/\text{s}$ at the equator scaled proportionally to the cell area to the power $3/2$. The background values of vertical viscosity and diffusivity are 10^{-4} and $10^{-5} \text{ m}^2/\text{s}$, while the shear-induced vertical mixing is parameterized by the Munk-Anderson scheme with maximum values of 10^{-2} and $10^{-3} \text{ m}^2/\text{s}$, respectively. The time step is 4 minutes, and the barotropic velocities are calculated in a nested cycle with a step of 5 seconds. For the calculation of the momentum transport, the centered difference scheme is applied, and for the tracers we use the flux-corrected transport scheme [12].

The computations were carried out for 5 years of model time, from 1978 to 1982. The atmosphere-ocean boundary fluxes were calculated from the array of atmospheric properties and the bulk-formulas of the CORE-II protocol [13]. The model starts from zero velocities and the annual mean temperature and salinity fields of WOA09 [14,15]. The solution is saved as consecutive 5-day means. As in [8], in this paper we analyze the mean characteristics for 1979 – 1982.

2.3. Eddy MHT calculation

For calculation of the eddy heat transport through an xz -section, ocean general circulation models usually apply a formula of the following kind:

$$Q_E = \iint \rho C_p \theta V dz dx - \iint \rho C_p \langle \theta \rangle \langle V \rangle dz dx - \iint A_H \rho C_p \frac{d\theta}{dy} dz dx \quad (1)$$

Here V is the meridional velocity component, θ is the potential temperature, ρ is the density and C_p is the heat capacity of the ocean water; A_H denotes the coefficient of model diffusion of tracers, which can depend on the grid parameters and properties of the solution.

The first two terms in (1) represent the difference of the total transport and its part due to the time-mean current. By definition, this difference is considered as explicitly resolved eddy heat transport. Angular brackets here denote averaging over the characteristic time scale of mesoscale eddies, which is on the order of 1 to 3 months [16] (in this paper we use the centered running mean over 90 days). The last term corresponds to the subgrid heat flux caused by artificial model tracer diffusion. Nowadays, the common practice in detailed studies of ocean circulation is the intention to make the contribution of diffusion terms very small, thus reproducing as much as possible the spectrum of ocean turbulence outside the boundary layers. In our experiment, this subgrid contribution to the eddy MHT was examined and turned out to be negligible.

It should be noted that this approach takes into account not only the actual eddy effects, but also all other short-period elements of the ocean circulation (“transients”), such as meandering of currents, TIWs, and rapid variability of the Ekman layer. Their contribution to the MHT is caused by the correlations of rapidly varying fields of temperature and velocity. Traditionally, in the literature they are all denoted by the term “eddy transport”, and in this article we will also keep this terminology.

3. Tropical instability waves and equatorial convergence of the eddy MHT

The heat balance of the Pacific equatorial region shows a significant contribution of the eddy heat flux convergence, which is associated with TIWs and partly compensates for the Ekman transport diverging from the equator. TIWs are perturbations of the temperature front near the ocean surface on either side of the well-known “cold tongue” of the East Pacific waters. They develop as a consequence of shear instability at the flanks of the South Equatorial Current, demonstrating pronounced temperature variations with magnitudes of 1-2°C, periods of 20-40 days, and wavelengths of 1000-2000 km. Typically, TIWs are formed around 100 – 110°W and spread westward, expanding in the north-south direction in case of weakening background temperature gradients [17,18].

Figure 1 shows the meridional distribution of the eddy MHT convergence, which was calculated in the band from 0.125°S to 0.125°N. It is compared with the simulation data [4] and the estimate [19] based on moored observations which were performed in 1979 – 1981.

It can be seen that the average values of the model eddy MHT convergence in the 110°W – 140°W interval almost coincide for the INMIO model (177 W/m²) and for model [4] (190 W/m²). The average value for the INMIO in the 110°W – 152°W interval (158 W/m²) turned out to be near the lower error estimation boundary of a similar average from [19] (245 ± 84 W/m²). To an appreciable degree, this is due to the divergence of the eddy heat fluxes from the equator near 141°W (-54 W/m²), where these fluxes have a component co-directed with the temperature gradient (see Figure 6 in [8]). If we exclude this outbreak and also a weaker one at 135°W, our results also agree quantitatively well with a spread of 100 – 350 W/m² obtained in the coupled simulation [20] for the upper 250-meter layer in the 160°E– 100°W band.

Within the error interval of the estimate [19] (95 ± 100 W/m²) are the INMIO data (164 W/m²) on the meridian 152°W. At 110°W, the model result (170 W/m²) appears significantly lower than that of [19] (380 ± 135 W/m²). Taking into account that the model [4] in this location showed results qualitatively close to the INMIO (as can be seen from their mean 190 W/m² and the distribution in their Fig. 8a), while the drifter-based estimate [21] suggests a somewhat higher value (180 W/m² only in the upper 50-meter layer, for the 105°W – 120°W band), it can be assumed that this difference between the models and observations in the region under consideration is systematic. Next, consider the thermohydrodynamic conditions in which it is formed.

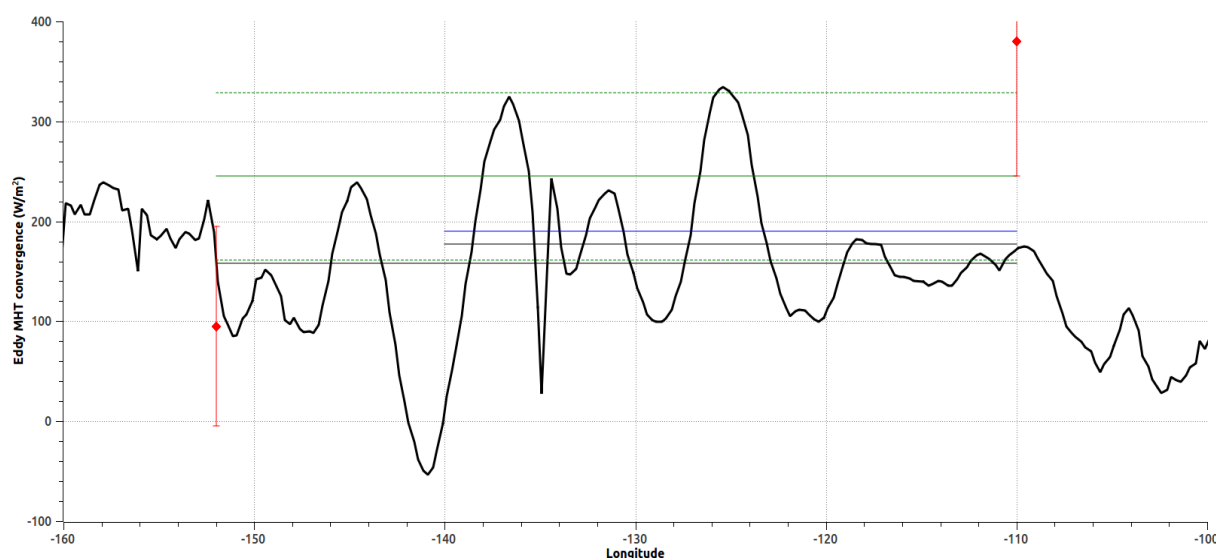


Figure 1. Equatorial distribution of the eddy heat flux convergence (W/m^2). Thick black curve denotes INMIO WOM-0.1 data. The solid blue and green lines are the average values of [4] and [19], and the thin black lines are the average INMIO values in respective longitude ranges. The dotted green lines mark the error estimation of the mean [19]. The red diamonds with error bars denote the data of [19] for the boundaries of the $110^\circ\text{W} - 152^\circ\text{W}$ interval.

4. Model solution in the Equatorial Pacific

4.1. Surface characteristics

To illustrate the circulation in the tropics of the Pacific Ocean, in Figure 2 we give the mean model SST and wind stress for 1979 – 1982 in the band from 2°S to 2°N . To validate these results, we use the temperature data from the global HadISST1 analysis [22] averaged for the same years. In our experiment, the wind stress, which is directly assimilated by the dynamical core of the INMIO model, is calculated based on the bulk-formulas and atmospheric CORE-II data [13] derived from the NCEP/NCAR reanalysis. For an independent comparison on this parameter, we use wind stress data of the ERA-Interim reanalysis.

An essential discrepancy between the model and observational data is the westward shift of the equatorial temperature minimum. This is a robust feature that persists when performing an experiment with a lower resolution (0.25°) under the same principal conditions (CORE-II protocol for the same years, the model with diffusion and biharmonic viscosity). It is almost unchanged in case of continuing the calculation for the next nine years (green curves in Figure 2(a)).

A comparison with Figure 1 shows that the zone of high eddy activity in the model extends to the west from 110°W , which roughly coincides with the zone of low SST and increased wind stress. However, microwave sounding data (which appeared in later years, cited in [17]) indicate that a high level of mesoscale variability should begin farther to the east, approximately from the Galapagos Islands (90°W).

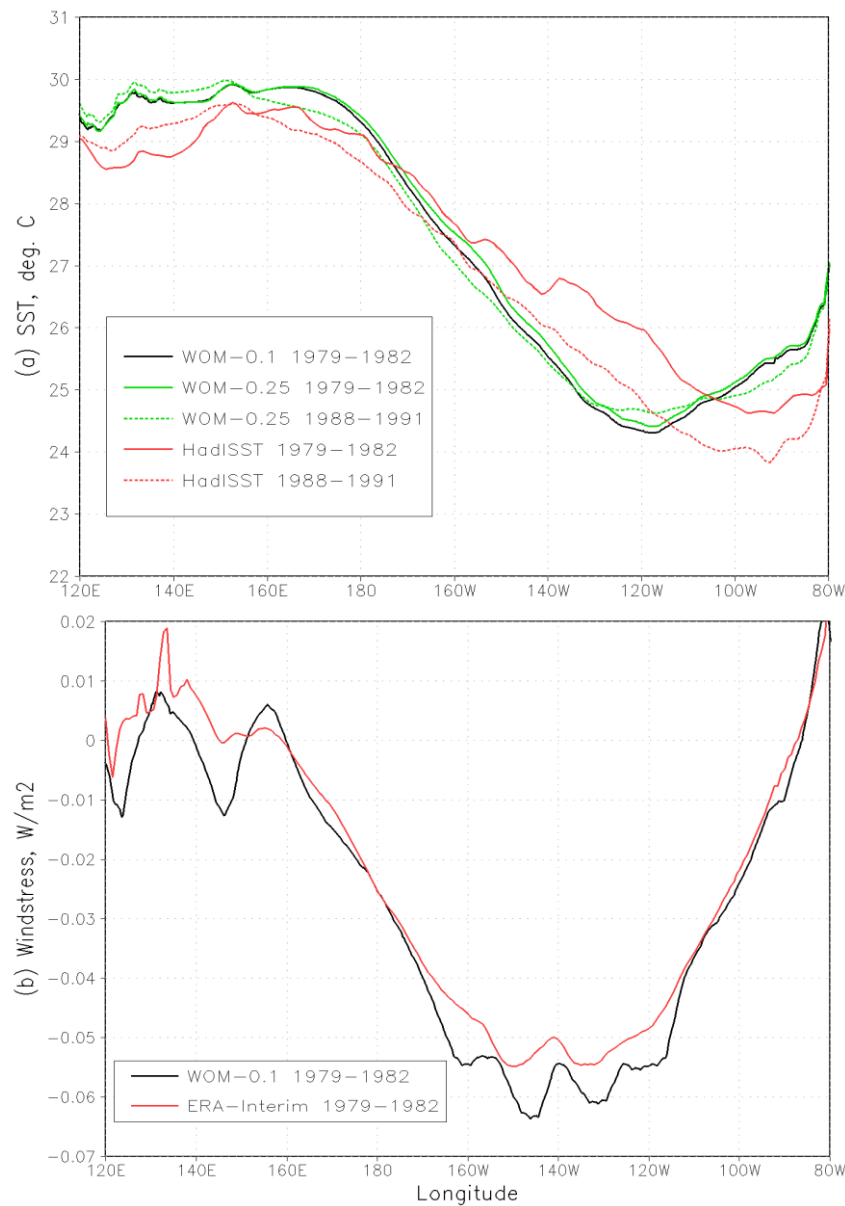


Figure 2. Sea surface temperature (a) and wind stress (b) averaged over specified years and in the 2°S – 2°N band.

4.2. Upper ocean temperatures

For a more detailed study of the SST bias, see in Figure 3 the vertical equatorial section for the model temperatures and for the WOA09 climatology data, in both cases averaged over the 2°S – 2°N band. From the mentioned 90°W meridian in the climatology data, when moving westward a gradual deepening of the thermocline begins. On the contrary, in the model data the thermocline proves to be substantially diffused and deepens abruptly starting from 110–120°W. Its correct simulation probably would require improvement of the numerical transport schemes in the model dynamical core and / or of the vertical mixing parameterization. Also, an increased value of the eastern wind stress can contribute to the rise of cold waters at longitudes 120–160°W through the Ekman upwelling mechanism. Therefore, the accuracy of setting of atmospheric conditions and the correctness of the

particular bulk formulas for the transformation of ocean and atmosphere parameters into heat, mass, and momentum fluxes deserve a separate consideration for this region.

Figure 3(c) shows that the zone of cold bias of the model solution extends to the east to about 105°W. In the tropics, the upper ocean heat content balance is determined primarily by the divergence of oceanic heat fluxes, in contrast to the midlatitudes where the influence of the seasonal cycle of surface fluxes is more important (see Section 8 of [23] and references therein). Therefore, we can conclude that the cold bias in the upper layer of the model ocean may be caused by the lack of heat transport convergence by TIWs in the 105-170°W band. This agrees qualitatively with the data in Figure 1, where the average value for the INMIO model in the 110°W – 152°W interval lies near the lower error bound of the similar mean taken from the data of [19].

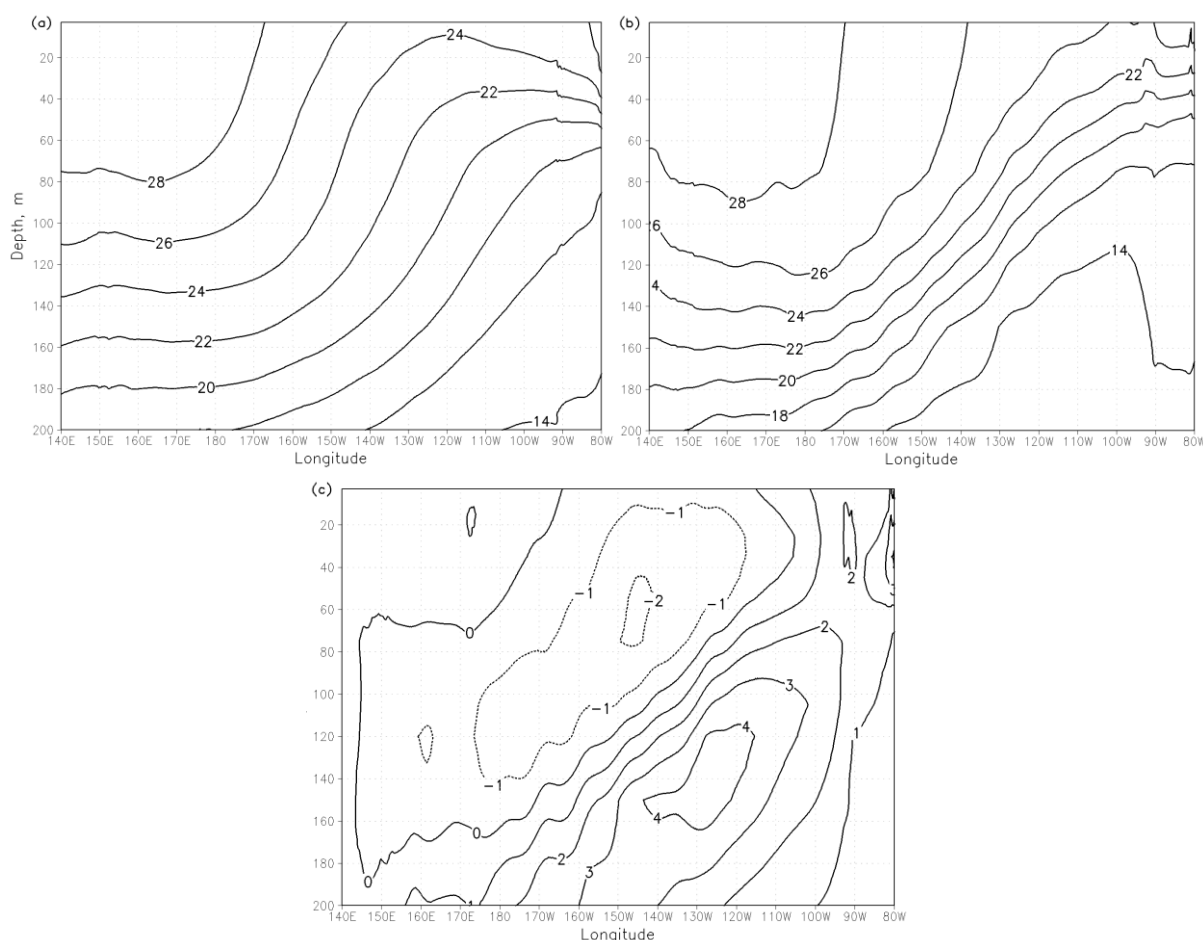


Figure 3. Longitude-depth cross-section of the temperature field (°C) averaged over the 2°S – 2°N band. INMIO WOM-0.1 model, 1979-1982 mean (a); annual mean WOA09 climatology (b); their difference (c).

Moreover, the example of Figure 4 for 110°W and 140°W shows that at these longitudes the isotherms of the model solution curve upward near the equator stronger than the WOA09 isotherms. The diffused thermocline is also manifested. It is known that one of the effects of the eddy heat transport convergence by TIWs is the heating of the cold tongue in the 100-200 m top layer of the ocean and the corresponding flattening of isotherms through the mechanism of baroclinic instability. In nature, this allows the equator to be reached by waters warmer than in the current model results. Therefore, this example also supports the hypothesis that in this region the intensity of the TIWs in the model solution is underestimated relative to that present in nature.

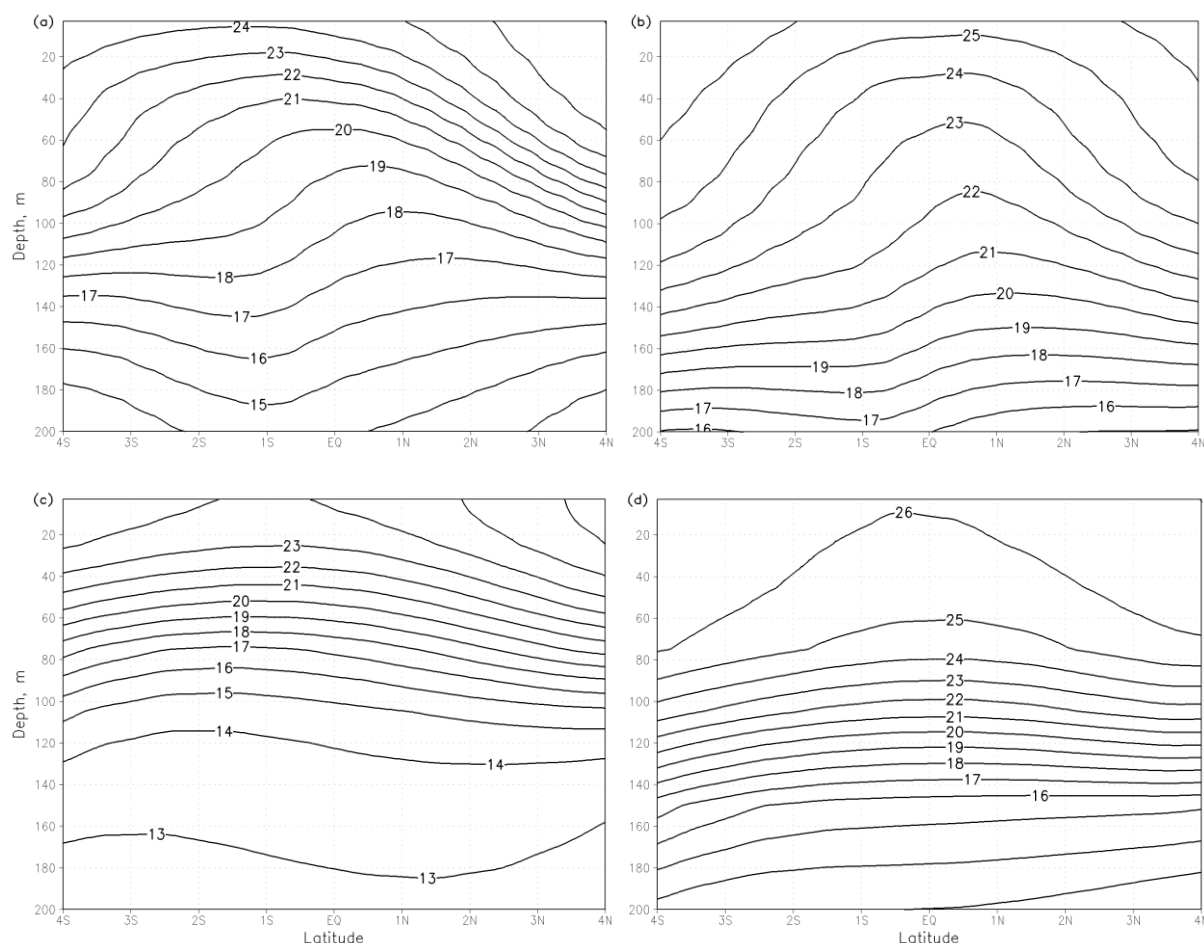


Figure 4. Latitude-depth cross-section of the temperature field (°C). INMIO WOM-0.1 model, 1979-1982 mean (a,b); annual mean WOA09 climatology (c,d); figures for 110°W (a,c); figures for 140°W (b,d).

5. Conclusions

We have studied some of the thermohydrodynamic characteristics of the Eastern Equatorial Pacific by means of the INMIO model with a 0.1° resolution and compared the results with reference data of observations and reanalysis. The main goal of this study was to investigate the conditions in which the equatorial convergence of eddy meridional heat transport is formed by tropical instability waves heating the cold tongue of the East Pacific waters. The model has a sufficient resolution to reproduce the TIWs, but near 110°W it showed significantly underestimated convergence values relative to estimates of direct observations. Consideration of the mean temperature distributions in comparison with the WOA09 climatology has shown that a diffused thermocline is formed in the model solution, which excessively sharply deepens while moving westward. At the same time, in a latitude-depth section the isotherms curve strongly upward near the equator, and the cold waters move eastward a longer distance than in nature. This is an evidence in favour of the assumption that the underestimation of the eddy MHT convergence is a bias of the model rather than an error of the MHT observational assessments. Since a similar underestimation was found [8] in the results of a model study [4], it seems reasonable to further investigate the numerical mechanisms responsible for the distortions of the model solution mean characteristics considered in this paper. This may, in particular, concern the transport schemes or the parameterizations of vertical turbulent mixing.

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